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WEATHERED ZONES AND GLACIAL CHRONOLOGY IN SOUTHERN INDIANA

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ABSTRACT

The Illinoian drift of southern Indiana has been leached of its calcium carbonate to an average depth of 158 inches. The depth of leaching in the Wisconsin drift varies from about 4½ feet along the Shelbyville moraine to 3 feet along the Bloomington moraine. Fifty-nine localities were found where calcareous Wisconsin drift, loess, or lake silt rests upon leached Illinoian drift. Measurements at these localities indicate an average leaching of 6 feet during the Sangamon interglacial stage. From 4 to 5 feet of gumbotil were formed upon the Illinoian drift during Sangamon time. This compares closely with the amount of gumbotil formed to the west of Indiana, but leaching did not penetrate to so great a depth in Indiana as elsewhere. This may have been due to lower relief and higher water table in Indiana during Sangamon time than in the states west of Indiana.

Comparison of the weathering of the silts in southwestern Indiana shows the presence there of local loess and lake silt of Illinoian age and extensive loess and lake silt of Wisconsin age. The main body of loess in Indiana is the equivalent of the Peorian loess to the west. Within the Tazewell drift area two loesses occur which are the equivalents of the Tazewell and Iowan loesses of Illinois. Extensive lake silt of Wisconsin age occurs along the tributaries of the Wabash and Ohio rivers. The ponding of these tributaries was due to the deposition of an extensive valley train along the lower Wabash, while it acted as a sluiceway for the Wisconsin ice sheet.

An estimate of 135,000 years is made for the duration of the Sangamon interglacial stage in southern Indiana. The Tazewell drift in Indiana has probably been undergoing leaching for at least 45,000 years.

INTRODUCTION

Leighton and MacClintock,[1] describing the soil profiles developed upon the Illinoian drift of southern Illinois, state that wherever the

[1] M. M. Leighton and Paul MacClintock, "Weathered Zones of the Drift Sheets of Illinois," *Jour. Geol.*, Vol. XXXVIII (1930), pp. 28–53.

Illinoian or older drifts are exposed in deep-enough sections five fairly easily recognizable horizons are discernible. These are:

Horizon 1. The surficial soil
Horizon 2. Chemically decomposed till, composed chiefly of alteration products and resistant constituents of the original till, and strikingly unlike the original till
Horizon 3. Leached and oxidized till, otherwise but little altered
Horizon 4. Oxidized till, but unleached and otherwise unaltered
Horizon 5. Unaltered till

Since these investigators have described these horizons and pointed out the variations in their characteristics under varying topographic conditions, it is not necessary to discuss their development in this paper. The present objective is to compare the soil profiles developed upon the Illinoian and Wisconsin drifts of Indiana with those in Illinois and other states west of Indiana and to describe the general stratigraphy of the glacial materials of southern Indiana. The field work consisted of the study of several hundred exposures of glacial materials and the making of several hundred borings with a soil auger to determine the character of the various soil horizons and particularly the depth of leaching. The area covered was roughly the southern half of Indiana.

SOIL PROFILES IN GLACIAL DRIFT IN INDIANA

The weathered zone in the Illinoian drift in Indiana has reached a mature stage of soil development and shows clearly recognizable subdivisions of the soil profile. Weathering of the Wisconsin drift has not yet reached a sufficiently mature stage for all soil horizons to be easily differentiated. The zone of secondary iron accumulation which is usually found near the boundary between Horizons 2 and 3 is not always present, and this makes the demarcation of their boundary somewhat difficult. Horizon 2, which in the Illinoian drift has reached the gumbo stage or its equivalent, in the Wisconsin drift still retains the original characteristics except for the leaching and oxidation which have occurred. However, there is little difficulty in distinguishing Horizons 4 and 5 in either the Illinoian or Wisconsin drift.

Figure 1 represents diagrammatically the soil profiles in the Il-

linoian and Wisconsin tills in southern Indiana. It may be noted that the three upper horizons in the Wisconsin till are only about one-third as thick as they are in the Illinoian till. The Illinoian till has a particularly thick Horizon 2, the one in which gumbotil de-

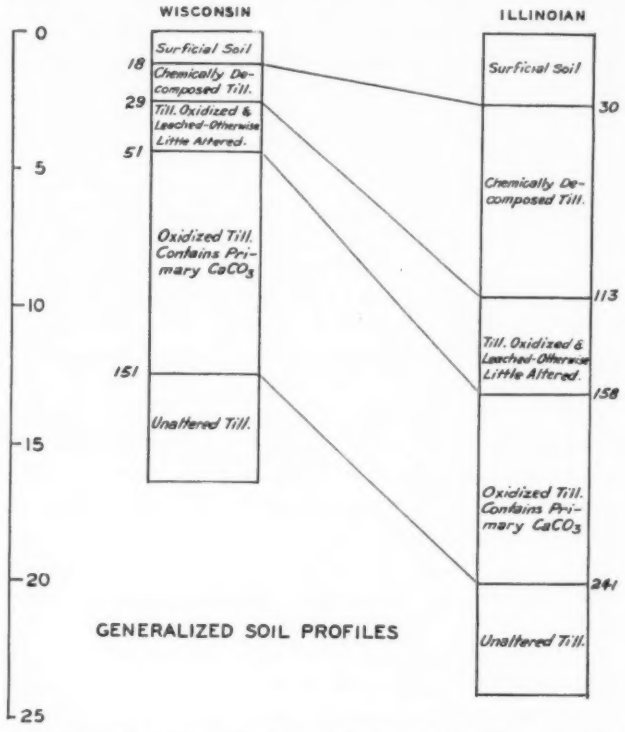

FIG. 1.—Generalized soil profiles for the Illinoian and Wisconsin tills of southern Indiana.

velops. The Wisconsin till is leached of its calcium carbonate to an average depth of 51 inches, whereas the Illinoian till has been leached to a depth of 158 inches. Horizon 4, which has undergone oxidation but not leaching, is thicker in the Wisconsin drift than in the Illinoian. This indicates, as pointed out by Kay and others, that leaching and oxidation go on more slowly as greater depth is attained

and the other weathering processes (chemical decomposition, disassociation, etc.) gradually catch up with the leaching and oxidation.

MAPPING OF THE ILLINOIAN-WISCONSIN BOUNDARY IN INDIANA

The notable difference between the Wisconsin and Illinoian soil profiles can be used for accurate mapping of the boundary between the two drifts. This boundary, as indicated by the difference in the depths to carbonates, is shown in Figure 2. Although in most of Indiana topographic contrasts between the Illinoian and Wisconsin till plains are so pronounced that their contact can be mapped rather accurately upon a topographic basis, the method used here permits of somewhat more detailed mapping of this boundary. Figures on the map with a minus sign (−) represent locations where carbonates were not reached because bedrock or boulders stopped the soil auger, but in many cases sufficient depth had been attained to leave no doubt as to whether the hole was in Illinoian or Wisconsin till.

DIFFERENTIATION OF THE SUBSTAGES OF THE WISCONSIN

An attempt was made to differentiate the tills of the substages[2] of the Wisconsin in Indiana by using the depth of leaching as a means of mapping the tills. This was attempted only with respect to the Tazewell and Cary substages, as the field work did not extend into the area of the Mankato drift. Using an average of several borings, the results are satisfactory enough; but when individual holes are compared, the results do not seem to permit much more accurate mapping of the substages than where they are mapped by moraines. The Tazewell drift can be easily distinguished from the Mankato, but there is some difficulty in differentiating the Tazewell from the Cary.

[2] In all earlier work on the Pleistocene of Indiana the Wisconsin drift has been subdivided into early, middle, and late Wisconsin. The portion of the Wisconsin drift covered in this paper is essentially that from the Shelbyville to the Bloomington moraine, heretofore designated as "early Wisconsin." It seems advisable to accept the proposed classification of Kay and Leighton which subdivides the Wisconsin into four substages—Iowan, Tazewell, Cary, and Mankato. What has heretofore been called "early Wisconsin" in Indiana will be designated as "Tazewell" in the present paper. Obviously, the term "early Wisconsin" is no longer good, for the Iowan is early Wisconsin.

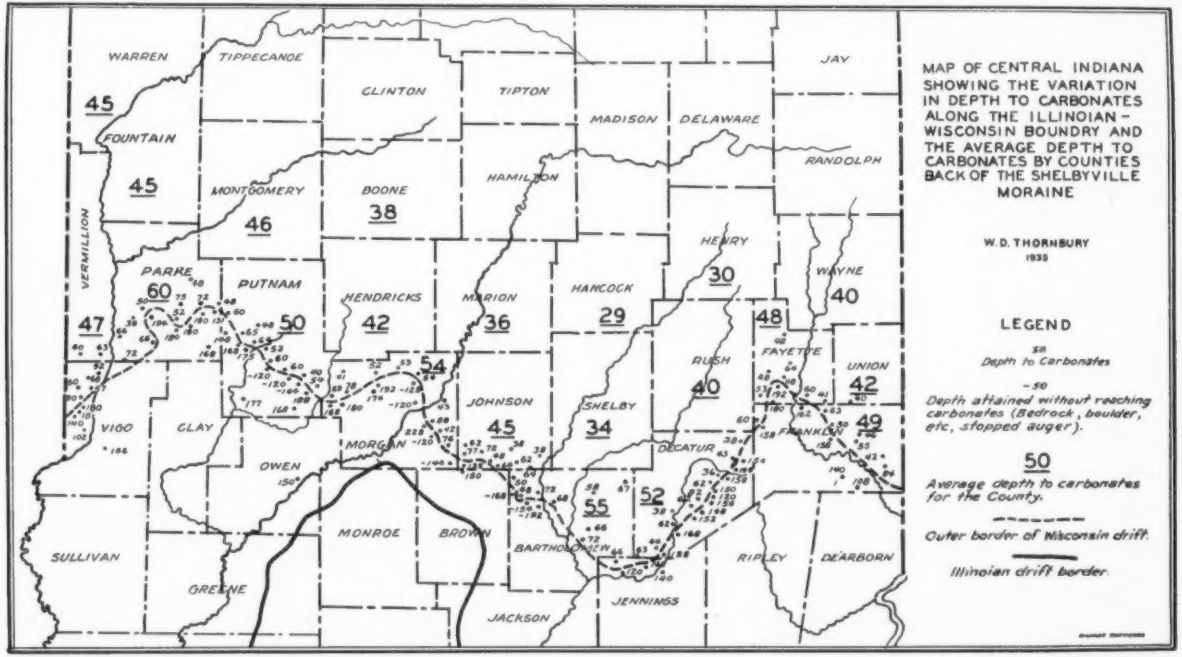

FIG. 2.—Map of central Indiana showing the variation in depth to carbonates along the Illinoian-Wisconsin boundary and the average depth to carbonates by counties back of the Shelbyville moraine.

The time interval between the Tazewell and Cary substages was probably about 10,000 years. This would probably make a difference of about 1 foot in the depth of leaching. Where the difference in the depth of leaching is only a matter of inches, there are several other factors which may cause as much variation in the depth to carbonates as the time factor. Chief of these factors are the texture of the till, the composition of the till, the height of the water table, and the variation in topographic conditions.

In spite of the somewhat unsatisfactory results obtained, they do show that, where the average of a large number of borings is used, there is a progressive decrease in the depth to carbonates as one goes

TABLE 1

	Depth to Carbonates in Inches
64 holes immediately back of Shelbyville moraine	58
181 holes between Shelbyville and Champaign moraines	51
111 holes between Champaign and Bloomington moraines	42
100 holes immediately back of Bloomington moraine	38

northward from the Shelbyville moraine. This is shown in Table 1, which gives the average depth to carbonates going northward from the Shelbyville moraine through the area of Tazewell drift to that of the Cary drift. The moraines are as mapped by Leverett.

The borings were made under as similar topographic conditions as possible. Figure 2 gives the average depth to carbonates by counties north of the Shelbyville moraine; and it also shows a gradual decrease in the depth of leaching northward, although there is considerable variation between counties, owing to differences in the topography and geology. The till is much more calcareous in the counties having extensive underlying limestone than in those in which sandstones and shales predominate. The bedrock affects the depth of leaching, as does the better drainage in counties crossed by major streams.

SEARCH FOR PRE-ILLINOIAN DRIFT

So far, no drift older than Illinoian has been discovered in Indiana. Pre-Illinoian drift, however, has been recognized in Illinois close to the Indiana state line by MacClintock,[3] who has described

[3] Paul MacClintock, "Recent Discoveries of Pre-Illinoian Drift in Southern Illinois," *Ill. State Geol. Surv., Rept. Investigations No. 19* (1929), Part II, pp. 27–57.

several exposures in Clark County, Illinois, just west of Vigo County, Indiana. Hopes were entertained of finding exposures of pre-Illinoian drift in Indiana in either Vigo or Clay counties; but in spite of careful search, no such exposures were discovered. The failure to find pre-Illinoian drift in Indiana may be due to its absence or to lack of sufficient relief. The relief along the tributaries of the Wabash on the Indiana side is not nearly so great as along the streams which enter the Wabash from the Illinois side.

BURIED ILLINOIAN SOILS AND WEATHERED ZONES

Leverett has described numerous occurrences in Indiana of a buried Illinoian soil or drift beneath Wisconsin drift. Much of his information was from well records. Such records are more subject to inaccuracies of interpretation than surface exposures. Consequently, detailed search was made for actual exposures of Wisconsin till or other glacial deposits resting upon weathered Illinoian drift. Reliance was placed only on a study of these. In a few cases, where only part of the weathered Illinoian material was exposed, a soil auger was used to determine the complete thickness of the Illinoian leached zone.

Within the portion of Indiana extending approximately 20 miles north of the outer border of the Tazewell drift forty-six localities were found where calcareous Wisconsin drift could be seen resting upon noncalcareous Illinoian drift. In addition to these forty-six localities within the area of the Tazewell drift, thirteen localities were found south of the Shelbyville moraine where buried Illinoian till, lake silt, or gravel showing a leached zone occurs beneath Wisconsin loess or lake silt. More exposures were found in Decatur and Putnam counties (see Figs. 2 and 3) than elsewhere, as the relief in these two counties is such as to preserve bluffs of unconsolidated glacial deposits. The places where weathered Illinoian drift was found beneath Wisconsin loess or lake silt are in the Wabash Valley and the adjacent uplands extending from Vigo County, at the north, to Posey County, at the extreme southwest tip of Indiana.

SECTIONS SHOWING BURIED ILLINOIAN DRIFT

A few representative sections will be given to show the amount of leaching of the Wisconsin and Illinoian drifts and the thickness of

the weathered zone in the buried Illinioan drift. Figure 3 gives in columnar form twenty selected sections from the Tazewell area north of the Shelbyville moraine. The numbers of the sections described below correspond to the numbers in Figure 3.

SEC. 63. DECATUR COUNTY, STREAM BLUFF
SE. ¼, SW. ¼, Sec. 5, T. 11 N., R. 10 W.

	Thickness in Feet
Wisconsin:	
Till, brown clayey, leached	5
Till, brown pebbly, oxidized, calcareous	8–9
Illinoian:	
Silt, gray, medium-fine, leached	1
Gumbotil, brownish, cherty, leached	1–1½
Till, red, cherty, leached	2½–3
Till, red, cherty, calcareous	1

SEC. 75. BARTHOLOMEW COUNTY, ROAD CUT
Near SW. Cor., Sec. 33, T. 9 N., R. 6 E.

	Thickness in Feet
Wisconsin:	
Sand, leached	3
Silt, leached	3
Till, brown, blocky, oxidized, calcareous	9–10
Till, bluish-gray, unweathered	3
Silt, brown, contains organic matter, calcareous	1½
Illinoian:	
Gumbosand, brownish-gray, leached	5

The foregoing is a rather interesting section, as it contains leached sands of both Wisconsin and Illinoian ages. It is at the edge of the

SEC. 83. PUTNAM COUNTY, STREAM BLUFF
NE. ¼, NE. ¼, Sec. 14, T. 16 N., R. 3 W.

	Thickness in Feet
Wisconsin:	
Till, fine-textured, leached	4
Till, brownish-gray, pebbly, oxidized, calcareous	7–8
Till, bluish-gray, unweathered	12–15
Illinoian:	
Gumbotil, blocky, contains organic material, leached	3
Till, brownish-green, silty, leached	3
Till, bluish-gray, calcareous	1–3

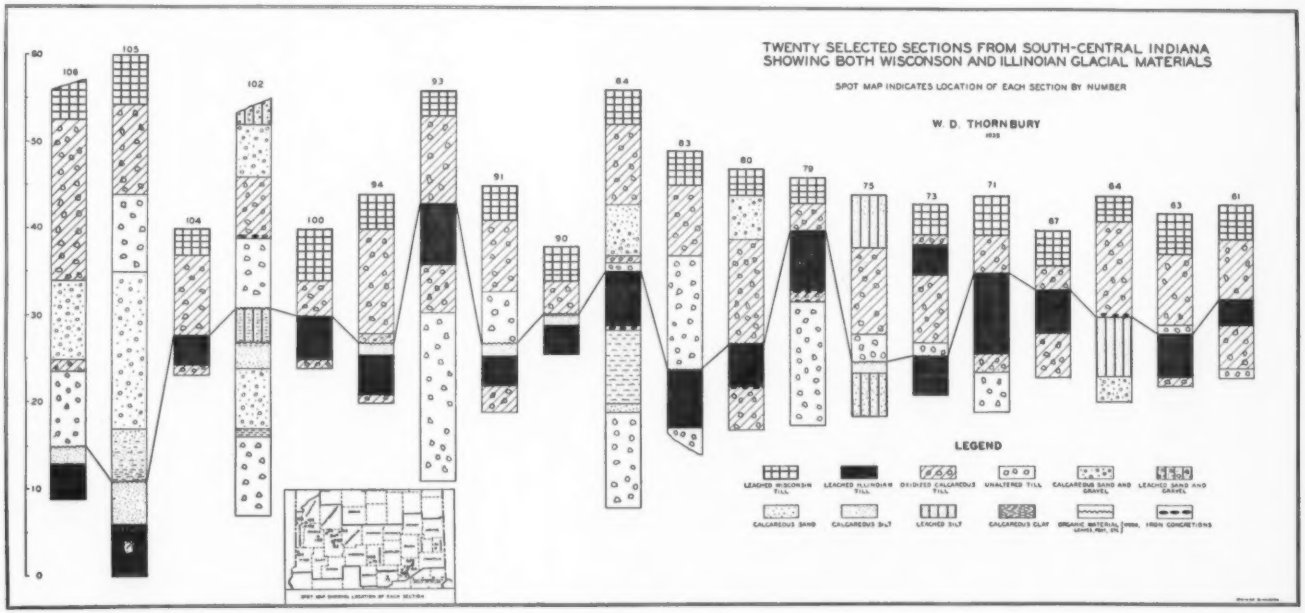
TWENTY SELECTED SECTIONS FROM SOUTH-CENTRAL INDIANA
SHOWING BOTH WISCONSON AND ILLINOIAN GLACIAL MATERIALS
SPOT MAP INDICATES LOCATION OF EACH SECTION BY NUMBER
W. D. THORNBURY
1938
LEGEND
LEACHED WISCONSIN TILL
LEACHED ILLINOIAN TILL
OXIDIZED CALCAREOUS TILL
UNALTERED TILL
CALCAREOUS SAND AND GRAVEL
LEACHED SAND AND GRAVEL
CALCAREOUS SAND
CALCAREOUS SILT
LEACHED SILT
CALCAREOUS CLAY
ORGANIC MATERIAL
IRON CONCRETIONS
106
105
104
102
100
94
93
91
90
84
83
80
79
75
73
71
67
64
63
61
50
40
30
20
10
0

FIG. 3

valley of the East Fork of White River, which acted as a sluiceway for Wisconsin melt waters. Numerous sand dunes of Wisconsin age occur along this valley.

The organic matter in the upper part of the Illinoian weathered zone in Section 83 suggests that very little erosion of that till occurred and that the 6–7 feet of noncalcareous till represent approximately the full amount of leaching during Sangamon time.

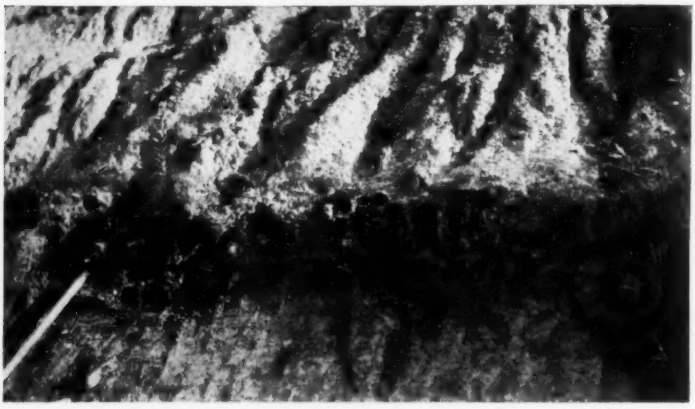

FIG. 4.—View of portion of stream bluff in Putnam County, Indiana, showing calcareous Wisconsin drift resting upon peaty soil and gumbotil of Illinoian age. (See sec. 94 for description.)

SEC. 94. PUTNAM COUNTY, STREAM BLUFF

NW. $\frac{1}{4}$, NW. $\frac{1}{4}$, Sec. 29, T. 16 N., R. 5 W.

	Thickness in Feet
Wisconsin:	
Till, silty, yellowish-brown, leached	4
Till, pebbly, oxidized, calcareous	12–13
Illinoian:	
Silt, large amount of organic material, leached	1–1$\frac{1}{2}$
Gumbotil, gray, silty, leached	4$\frac{1}{2}$

Only part of the gumbotil is exposed in Section 94. A soil auger was used to determine the depth to which leaching extends. Figure 4 is a partial view of this section.

AMOUNT OF LEACHING DURING SANGAMON TIME

In many of the above-mentioned forty-six sections part of the leached Illinoian drift obviously had been removed by the Wisconsin ice sheet. To take the average of the thicknesses of the Illinoian leached zone without regard to possible erosion would, of course, give too low a figure for the leaching during Sangamon time. In seventeen of the sections organic material, or what had the appearance of an old soil, is present at the top of the Illinoian drift. This would seem to indicate little or no erosion by the Wisconsin ice sheet. The average thickness of the leached Illinoian drift in these seventeen sections was found to be 64.2 inches, or nearly 5½ feet. This seems to be a fairly accurate estimate of the amount of leaching of the Illinoian drift during the Sangamon interglacial stage in this portion of Indiana.

Even more accurate estimates are obtainable from the Illinoian area south of the Tazewell drift. Along the lower Wabash Valley leached Illinoian drift may be seen at many places beneath loess or lake silt of Wisconsin age. Some erosion of the Illinoian drift likely occurred during Sangamon time; but as the relief in the Wabash Valley was very slight, the erosion was probably not great. At least it seems likely that the leached zone is more nearly complete in this area than within the area of the Tazewell drift, where considerable erosion by the Wisconsin ice sheet occurred.

LOESS AND LAKE SILTS OF SOUTHERN INDIANA

It is not the purpose of this paper to discuss in detail the loess and lake silts of the lower Wabash Valley; but since they bear to a considerable degree upon the problem under discussion, it may be appropriate to present a brief statement regarding them.

Loess was described in Indiana, near New Harmony, by Richard Owen as early as 1862. In 1903 Fuller and Clapp[4] proposed a twofold classification of the silts of southwestern Indiana. They proposed the name "marl-loess" for the silts occurring below an elevation of 500 feet and the term "common loess" for the silts above that elevation. They considered the marl-loess to be a fluvio-lacustrine

[4] M. L. Fuller and F. G. Clapp, "Marl-Loess of the Lower Wabash Valley," *Bull. Geol. Soc. Amer.*, Vol. XIV (1903), pp. 153–76.

material deposited in a large lake, but they were vague as to the conditions responsible for the formation of the lake. Shaw,[5] in 1915, objected to calling the "marl-loess" of Fuller and Clapp loess and argued that it was part of a great system of valley fillings formed in ponded tributaries of the Mississippi by the great volume of water carried by the main stream during the last glacial stage. He maintained that the term "loess" should be applied only to those silts which can definitely be shown to be of eolian origin.

The writer is essentially in agreement with Shaw's contentions. Much of the silt of southwestern Indiana is to be classed as lake silt rather than as true wind-blown loess. It is not difficult to distinguish the lake silts from the loess, because the lake silts are restricted to certain topographic levels which coincide with the lacustrine plains. However, this method is not applicable to buried silts, and it is very likely that some of the so-called "buried loesses" may be lake silts rather than loess. There is, therefore, need for detailed studies of the soil profiles developed in wind-blown silts and water-laid silts for other distinguishing criteria.

AGE OF THE LOESS AND LAKE SILTS

In southwestern Indiana, lake silts of two ages may be distinguished. Along the southeast margin of the Illinoian ice border there developed a series of marginal lakes, caused by the ponding of the southwestward drainage of that part of the state by the movement of the ice sheet across the valleys. Lakes Patoka, Cynthiana, and Pigeon are examples. These Illinoian lake plains have been described in the Patoka and Ditney folios of the United States Geological Survey. A second and more extensive system of lakes was formed during the Wisconsin glacial stage, owing to the ponding of the Ohio River and its tributaries and the tributaries of the Wabash River by the melt waters and outwash which the Wabash River carried while it acted as a sluiceway. The silts and clays deposited in the Wisconsin lakes occur at altitudes coincident with the downstream slope of the valley train developed in the Wabash Valley. These altitudes, along the Wabash Valley, vary from about 460 feet just south of the

[5] E. W. Shaw, "On the Origin of the Loess of Southwestern Indiana," *Science*, Vol. XLI (new ser., 1915), pp. 104–8.

Shelbyville moraine to about 400 feet at the junction of the Wabash and Ohio rivers.

The difference in the ages of the silts of these two lake systems is shown by the amount of leaching which they have undergone. The older silts, such as in Lake Patoka, are leached from 12 to 15 feet, which corresponds closely with the leaching of the Illinoian drift. The Wisconsin age of the other lake silts is shown by the fact that the average of twenty-one determinations of the depth to carbonates in them is 58 inches. This is only slightly greater than the depth to carbonates in the Tazewell drift, which is 51 inches. The slightly deeper leaching in the lake silts may be due to their looser texture, or it may mean that they are slightly older than the Tazewell drift. This point will be discussed later. But there can be no doubt that they are Wisconsin in age.

In addition to the two sets of lake silts in the lower Wabash Valley there are extensive deposits of wind-blown silt, or true loess. The loess is most abundant in the southwest corner of the state in the angle between the Wabash and Ohio rivers; but it occurs up the Wabash as far north as Warren and Fountain counties, which are within the Shelbyville moraine, and up the Ohio as far east as Perry County. Posey County, in the southwest corner of the state, is practically covered with a veneer of loess which in places attains a thickness of 40–50 feet.

Most of the loess of the Wabash and Ohio valleys is of Wisconsin age and is the equivalent of the Peorian loess as now defined by Kay and Leighton.[6] The term "Peorian" is now applied by the Iowa and Illinois Geological Surveys to the loess which in Iowa lies upon the Iowan drift and around its border, and to that in Illinois which lies above the late Sangamon loess outside the Tazewell drift. This loess is leached to an average depth of $5\frac{1}{2}$ feet in Iowa. Within the area of the Tazewell drift the Illinois Survey now recognizes two loesses: one (Iowan) beneath the Tazewell drift but above the late Sangamon loess, and the other (Tazewell) resting upon the Tazewell drift. Outside the Tazewell border the two cannot be distinguished and together constitute the Peorian loess.

[6] G. F. Kay and M. M. Leighton, "Eldoran Epoch of the Pleistocene Period," *Bull. Geol. Soc. Amer.*, Vol. XLIV (1933), pp. 669–74.

Most of the southern Indiana loess lies outside the Tazewell drift upon Illinoian drift or nonglacial materials. There is little evidence, in Indiana, of a late Sangamon loess such as has been recognized in Illinois. The only area in Indiana in which loess of Illinoian age seems to exist is around the margins of the Illinoian lake flats described above. In northern Dubois County and northeastern Pike County some fairly extensive deposits of silty material around the margin of the Lake Patoka plain seem to be of Illinoian age, as

FIG. 5.—Road cut in southern Gibson County, Indiana, showing undulating contact between leached and calcareous Peorian loess.

carbonates are not encountered in them at depths less than 12–15 feet. Small patches of similar materials occur in Gibson County. This loess appears to be more closely associated with the Illinoian glacial stage than with the Sangamon interglacial stage. However, most of the silts in the lower Wabash Valley are either lake silts or loess of Wisconsin age. Results of thirty leaching determinations along the Wabash Valley from the Shelbyville moraine to the southern end of the state show an average depth to carbonates of 64 inches in the loessial deposits not of Illinoian age. This clearly indicates that the main body of loess is of Wisconsin age and essentially the equivalent of the Peorian loess to the west. It further suggests that

loess accumulation may have begun in Indiana about the same time that it did in Iowa and Illinois, since the depth of leaching is only a few inches less than in the Peorian loess in Iowa. Since the silts composing the loess were derived from the flood plain of the Wabash while it carried drainage from the Wisconsin ice sheet, it appears likely that the Wabash carried considerable outwash as early as the Iowan substage. Figure 5 is a view of the Peorian loess which shows the undulating boundary between the leached and calcareous portions. This line stands out sharply because of the greater moisture-retaining capacity of the calcareous loess at the bottom of the road cut.

SECTIONS SHOWING LOESS OR LAKE SILT UPON LEACHED ILLINOIAN DRIFT

The area of Peorian loess south of the Tazewell drift is particularly favorable for determining the amount of leaching of the Illinoian drift during Sangamon time, because the veneer of loess or lake silt has protected the buried Illinoian weathered zone from subsequent weathering and erosion. It is probably true that the total amount of leaching during Sangamon time is not represented by the present thickness of the buried leached zone, since some subaerial erosion of this zone took place during Sangamon time; but it is likely that this zone is more nearly complete here than within the area of the Tazewell drift.

Figure 6 shows in columnar form twenty selected sections of loess and lake silts taken along the Wabash Valley from Warren County on the north to the Ohio River on the south. All the sections shown in Figure 6 except 1, 5, 7, and 8 occur south of the Tazewell drift, and in most cases they show a buried weathered zone including leached Illinoian till, gravel, sand, or silt. Following are brief descriptions of four of the sections. The numbers correspond to those in Figure 6.

Section 23 is interesting because it contains lake deposits of both Wisconsin and Illinoian age. The lowest silts and clays are deposits made in an Illinoian lake before the Illinoian ice sheet had advanced over the area. The leached sand above these silts and clays apparently represents outwash into this lake from the southward-advancing ice front. The presence of 3 feet of leached till indicates that

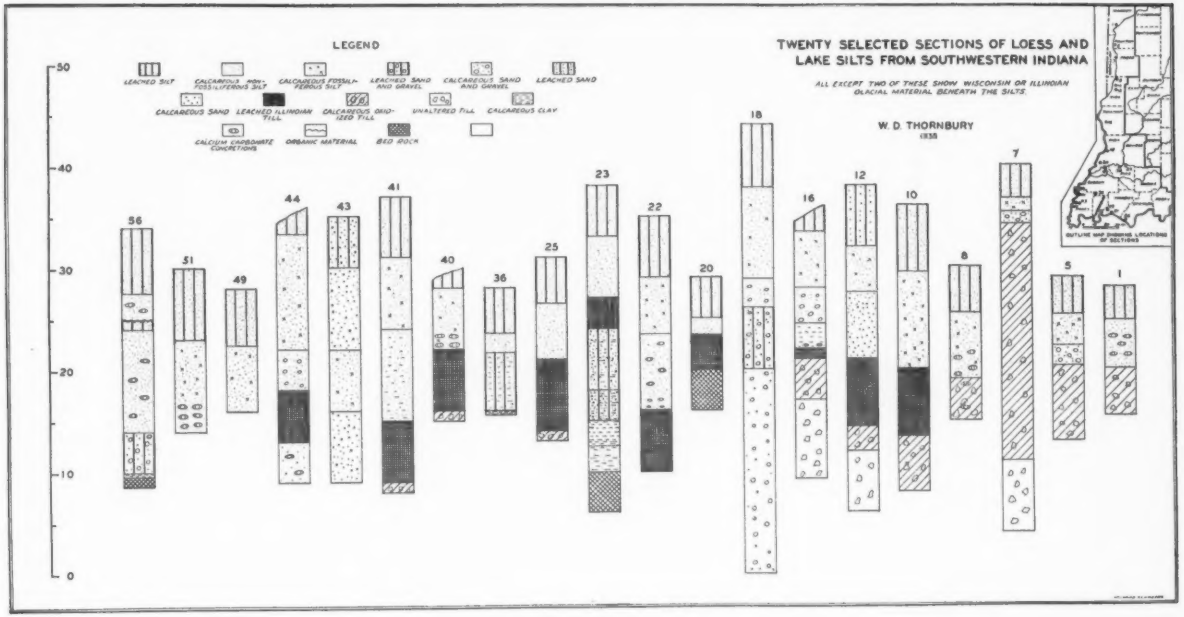
LEGEND
LEACHED SILT
CALCAREOUS NON-FOSSILIFEROUS SILT
CALCAREOUS FOSSILIFEROUS SILT
LEACHED SAND AND GRAVEL
CALCAREOUS SAND AND GRAVEL
LEACHED SAND
CALCAREOUS SAND
LEACHED ILLINOIAN TILL
CALCAREOUS ORD-ISIAN TILL
UNALTERED TILL
CALCAREOUS CLAY
CALCIUM CARBONATE CONCRETIONS
ORGANIC MATERIAL
BED ROCK
TWENTY SELECTED SECTIONS OF LOESS AND LAKE SILTS FROM SOUTHWESTERN INDIANA
ALL EXCEPT TWO OF THESE SHOW WISCONSIN OR ILLINOIAN GLACIAL MATERIAL BENEATH THE SILTS.
W. D. THORNBURY
1935
OUTLINE MAP SHOWING LOCATIONS OF SECTIONS
50
40
30
20
10
0
56
51
49
44
43
41
40
36
25
23
22
20
18
16
12
10
8
7
5
1

FIG. 6

the ice advanced temporarily onto the lake site. Following withdrawal of the ice, the lake deposits, outwash sand, and till were

SEC. 10. VIGO COUNTY, ROAD CUT
NE. ¼, NW. ¼, Sec. 6, T. 12 N., R. 8 W.

	Thickness in Feet
Wisconsin:	
Loess, brownish-buff, leached	6–6½
Loess, very fossiliferous, calcareous	9–10
Illinoian:	
Till, brownish-gray, leached	6–8
Till, brownish-gray, pebbly, oxidized, calcareous	3–8

SEC. 18. KNOX COUNTY, STREAM BLUFF ALONG WABASH RIVER A SHORT DISTANCE SOUTH OF THE SITE OF OLD FORT KNOX, ABOUT 2 MILES NORTHEAST OF VINCENNES

	Thickness in Feet
Wisconsin:	
Loess, brown leached	6
Loess, scattered fossils, calcareous	8–10
Silt, red, compact, pebbly, calcareous	2½–3
Illinoian:	
Gravels, deeply weathered, leached	5½–6
Gravels, cemented into conglomerate, calcareous	20

SEC. 23. PIKE COUNTY, ROAD CUT ON STATE ROAD NO. 61 AT BRIDGE OVER EAST FORK OF WHITE RIVER, NORTH OF PETERSBURG
Near SE. Cor., Sec. 16, T. 1 N., R. 8 W.

	Thickness in Feet
Wisconsin:	
Lake silt, reddish-brown, leached	4½–5
Lake silt, buff, fine texture, calcareous	5–6
Illinoian:	
Till, pebbly, leached	2½–3
Sand, stratified, interbedded clay, leached	5–6
Sand and clay, alternate sand and clay, leached	3
Clay, varved, compact, calcareous	2½–3
Silt, clayey, not varved, calcareous	2–3

leached to an aggregate depth of 10 feet before a Wisconsin lake covered the area. The greater than normal thickness of this leached horizon is probably due to the large amount of sand in it.

Figure 7 is a view of Section 36, which is located about 1½ miles east of the village of Cynthiana, from which Lake Cynthiana is named. This section is particularly interesting because it was formerly thought that the lacustrine plain which marks the site of Lake Cynthiana was composed solely of Illinoian silts. This section shows, however, that the upper 6 or 7 feet of silt are of Wisconsin age. The

SEC. 36. GIBSON COUNTY, EXPOSURE ALONG DRAINAGE DITCH LEADING INTO FLAT CREEK

Near SW. Cor., SE. ¼, Sec. 7, T. 4 S., R. 11 W.

	Thickness in Feet
Wisconsin:	
Lake silt, brown, leached	4–4½
Lake silt, mottled, loose texture, calcareous	1½–2
Illinoian:	
Lake silt, gumbo texture, iron concretions, leached	5½–6½
Lake silt, bluish-gray, compact, calcareous	1

following history is indicated for Lake Cynthiana. The water body was a marginal lake at the southeast edge of the Illinoian lobe, along with other similar lakes, such as Patoka and Pigeon. During its existence a considerable thickness of clays and silts was laid down over the floor of the lake. With the withdrawal of the Illinoian ice sheet the lake was drained, and these clays and silts were exposed as a lacustrine plain. During Sangamon time, these clays and silts were leached to a depth of about 6 feet. Very early in Wisconsin time (probably during the Iowan substage) another system of lakes formed along the tributaries of the Wabash River, owing to the ponding of these tributaries by the Wabash Valley train. One of these lakes extended over the site of the former lake, and in it were deposited several feet of silts and clays, burying the weathered Illinoian lake deposits. Since the drainage of the Wisconsin lake, leaching has reached to a depth slightly over 4 feet.

In all, thirteen sections were found exposing a buried Illinoian leached zone beneath Wisconsin loess or lake silt. In these the average thickness of this leached zone is 6 feet. This would seem representative for the Sangamon interglacial stage in southern Indiana.

This compares well with the average of 64.2 inches for Sangamon leaching in the area of the Tazewell drift.

COMPARISON OF INDIANA GLACIAL STRATIGRAPHY WITH THAT IN ILLINOIS

The Farm Creek Pleistocene section near Peoria, Illinois, is one of the most famous in the Middle West. A condensed description of

FIG. 7.—Side of drainage ditch east of Cynthiana, Gibson County, Indiana, showing Wisconsin lake silts upon Illinoian lake silts. (See sec. 36 for description.)

that section is given in Table 2 in order to compare the glacial stratigraphy in Indiana with that in Illinois. The designations of the various horizons according to early descriptions are given in the left-hand column; and the present interpretation of the section, as kindly

supplied the writer by Dr. M. M. Leighton, director of the Illinois Geological Survey, is given in the right-hand column.

Three loesses are present in the Farm Creek section. This section lies within the Tazewell drift border and contains a loess above and below the Tazewell drift. These two loesses are indistinguishable outside the Tazewell drift and together constitute the Peorian loess. Both are considered Wisconsin in age. The lowest loess is designated as "late Sangamon," and is associated with the Illinoian glaciation rather than with the Wisconsin.

TABLE 2

FARM CREEK PLEISTOCENE SECTION

OLD INTERPRETATION	PRESENT INTERPRETATION
7. Post-Bloomington loess	7. Tazewell loess
6. Bloomington gravels 5. Shelbyville till	6, 5. Tazewell gravels and till
4. Peorian loess	4. Iowan loess
3. Sangamon soil and loess	3. Late Sangamon soil and loess
2. Illinoian gumbotil	2. Illinoian mesotil
1. Illinoian till, unleached	1. Illinoian till, unleached

In Indiana most of the loess occurs outside the Tazewell drift area and thus corresponds to the Peorian loess. Loess does occur, however, within the Tazewell drift area. Sections 1, 5, 7, and 8 of Figure 6 are so located, and in each section the loess rests upon till which, though oxidized, has not been leached. This shows that the loess was deposited very shortly after the till. Loess can be seen at numerous localities on both sides of the Wabash River in Vigo, Parke, Vermillion, Fountain, and Warren counties. It has been traced as far north as Attica, in Fountain County. This loess is the equivalent of the Tazewell loess of the Farm Creek section, as now interpreted. That it is younger than the Peorian loess is shown by the fact that the seven sections studied show an average depth of leaching of 47 inches, as compared with 64 inches for the Peorian loess. That its deposition followed very closely upon the deposition of the Tazewell drift is further shown by the fact that the Tazewell drift, where not covered by loess in this area, is leached to an average depth of 46

inches. The very slightly greater depth of leaching in the loess can be explained by its looser texture.

The existence in Indiana of a buried loess to be correlated with the Iowan loess of the Farm Creek section is also suggested by three sections found containing buried silts beneath the Tazewell drift.

SECTION IN FOUNTAIN COUNTY, STREAM BLUFF ALONG MILL CREEK

NW. ¼, SE. ¼, Sec. 32, T. 18 N., R. 8 W.

	Thickness in Feet
Wisconsin:	
Till, leached	5
Till, oxidized, calcareous	10–11
Till, unweathered	8–9
Gravels, calcareous	18
Silt, brownish-gray, organic material in upper 1 foot, fossiliferous, calcareous	4–5
Illinoian (?):	
Mesotil (?) noncalcareous clay, possibly glacial	8–9

The fossiliferous silt near the bottom of the above-described section has all the appearances of a buried loess. Another section, described below, with a buried silt in it was found in Vermillion County.

SECTION IN VERMILLION COUNTY, STREAM BLUFF ALONG TRIBUTARY OF BIG VERMILLION CREEK

Near SW. Cor., SE. ¼, Sec. 30, T. 18 N., R. 10 W.

	Thickness in Feet
Wisconsin:	
Till, leached	3½
Till, calcareous	15–20
Sand and gravels, calcareous	8–10
Till, calcareous	9–10
Silt, brownish-gray, wood fragments in upper few inches, scattered pebbles, calcareous	2–3
Illinoian:	
Gumbotil, gray, leached	4

There might be some question as to whether the silt which lies upon the Illinoian gumbotil in the foregoing section is a true loessial

deposit, since pebbles occur in it. It may be a water-laid silt rather than wind blown.

SECTION IN VIGO COUNTY, STREAM BLUFF ALONG SUGAR CREEK

NE. ¼, SE. ¼, Sec. 22, T. 12 N., R. 10 W.

	Thickness in Feet
Wisconsin:	
Till, sandy to silty texture, leached	4
Till, pebbly, oxidized, calcareous	10–11
Silt, brownish-gray, organic material in upper 6 inches, fossiliferous, calcareous	5–5½
Till, unweathered	7–8

The buried silt in the above section has all of the characteristics of buried loess. No section was found containing both the Tazewell and Iowan loess. The Tazewell loess is well developed along the Wabash Valley, but not enough deep cuts or bluffs were found to determine how extensive the Iowan loess is. It does appear, however, that the recognition of two Wisconsin loesses in the Farm Creek section—one above and the other below the Tazewell drift—is confirmed from field evidence in Indiana.

COMPARISON OF LEACHING DURING SANGAMON AND POST-WISCONSIN TIMES

In Indiana approximately 6 feet of leaching took place during Sangamon time. Weathering had progressed to the stage of a mature soil profile. From 4 to 5 feet of gumbotil or mesotil had formed upon the Illinoian drift.

One cannot but note that the depth of leaching of the buried Illinoian drift is not much more than the depth of leaching in the Tazewell drift. The average depth of leaching in the Tazewell drift is 4½ feet. At first thought this might seem to indicate that the duration of Sangamon time was but little greater than post-Tazewell time. However, there are reasons for concluding that Sangamon time was of much longer duration than post-Tazewell time.

Kay,[7] using 25,000 years as the approximate length of post-Mankato time, estimates post-Iowan time at 2.2 times this, or 55,000 years, and Sangamon time at 4.8 times 25,000 years, or 120,000 years. His estimate of 55,000 years for post-Iowan time is based upon the comparative leaching of the Mankato and Iowan drifts. The Mankato drift is leached $2\frac{1}{2}$ feet, as compared with $5\frac{1}{2}$ feet in the Iowan drift. This assumes a uniform rate of leaching downward, which he recognizes is probably not true; so these estimates are minimum figures. On this basis post-Tazewell time in Indiana would be 1.8 times post-Mankato time, or 45,000 years, since the Tazewell drift is leached to an average depth of $4\frac{1}{2}$ feet. This estimate for post-Tazewell time corresponds closely with estimates made by Antevs.[8] He estimates the culmination of the Iowan substage at 70,000–65,000 years ago, the Tazewell substage at 45,000–40,000 years ago, the Cary substage at 36,000 years ago, and the Mankato substage at 27,000 years ago.

Kay arrives at his estimate of 120,000 years for Sangamon time by assuming an average amount of leaching during that time of 12 feet. Regarding leaching during Sangamon time he states:

> Good sections of Illinoian gravels which have the same topographic position as Illinoian gumbotil and which underwent weathering throughout Sangamon time have been difficult to find either in Illinois or in Iowa. A section in the vicinity of Sparta in southern Illinois shows more than 12 feet of leached Illinoian gravels below leached Peorian loess. Other sections in Illinois and Iowa are sufficiently satisfactory to warrant the judgment that the Illinoian upland gravels where they were not overlain by late Sangamon loess were leached as much as 12 feet before deposition of the Peorian loess.
>
> No sections of Iowan drift overlying the upland Illinoian gravels are known, but since the Peorian loess which overlies the leached Illinoian gravels is thought by the writer and some other glacial geologists, including Alden and Leighton, to be younger than the Iowan drift the leaching of the gravels may be interpreted to have taken place in Sangamon time.
>
> As already stated in this paper, the Iowan gravels where not overlain by Peorian loess or by Wisconsin drift have been leached about 5 feet 6 inches,

[7] G. F. Kay, "Classification and Duration of the Pleistocene," *Bull. Geol. Soc. Amer.*, Vol. XLII (1931), pp. 425–66.

[8] Ernst Antevs, "Climatic Variations during the Last Glaciation in North America," *Bull. Amer. Met. Soc.*, Vol. XIX, No. 5 (1939), pp. 172–76.

practically the same as the Iowan till, and the Peorian loess where it lies on Iowan drift and has not been overlain by Wisconsin drift also has been leached to about the same depth as the Iowan till. The Wisconsin upland gravels, as stated also, are leached to $2\frac{1}{2}$ feet.

Thus we see that the Iowan drift is leached to nearly as great a depth as the buried Illinoian drift in Indiana, although Kay's estimate of 12 feet of leaching during Sangamon time is considerably greater than that found in Indiana. The discrepancy between the comparative depth of leaching during Sangamon time in Indiana and Iowa and Illinois may find explanation in terms of differences in relief and height of water table in the two areas.

Two facts seem to indicate that the Illinoian drift plain in Indiana during Sangamon time was one of very low relief and poor drainage. One fact is that the thickness of the leached horizon is less than indicated in Iowa and Illinois. The other fact, which seems to suggest a very flat, poorly drained plain, is that the effects of oxidation do not extend very much farther down than the leaching. On the average, oxidation has extended only 4 or 5 feet lower than leaching, and in numerous sections only a foot or two farther down. In many of the sections observed, only a foot or two of oxidized but calcareous till separated the leached till above from the unweathered till below. This is in striking contrast to the conditions in the Tazewell drift. The total thickness of the weathered zone (zone of leaching plus zone of oxidation) in the buried Illinoian drift averages about $10\frac{1}{2}$ feet ($6 + 4\frac{1}{2}$ feet). In the Tazewell drift these two horizons average $13\frac{1}{2}$ feet ($4\frac{1}{2} + 9$ feet). Yet, 4–5 feet of gumbotil were formed in the Illinoian drift, and none was formed in the Tazewell drift. This not only indicates a much longer period of weathering for the Illinoian drift but also seems to indicate very little relief, poor drainage, and a high water table during Sangamon time, as compared with the conditions on the Tazewell drift plain.

The strongest evidence that Sangamon time was greatly longer than post-Tazewell time is that from 4 to 5 feet of gumbotil developed upon the Illinoian drift during this time, whereas no gumbotil has yet formed on the Tazewell drift. Although the thickness of the leached zone in the buried Illinoian drift cannot be correlated very closely with Kay's estimates of the amount of leaching during Sanga-

mon time, the amount of gumbotil formed in Indiana does coincide rather well with his estimate of 5 feet in Iowa and Illinois.

RATE OF FORMATION OF GUMBOTIL

No very accurate data are available as to the rate of formation of gumbotil; but it does not develop, as pointed out by Kay, until after leaching has gone on for approximately 50,000 years, since it is not found on the Iowan drift, which has been exposed to leaching for about 55,000 years. A few localities were observed in Indiana, where the Tazewell drift occurs upon very flat and poorly drained areas, where there appears to be forming what might be called an "incipient gumbo," as the weathered till has the gray, sticky, claylike texture associated with gumbo. Assuming a period of leaching of 45,000 years for the Tazewell drift, we arrive at a figure substantially the same as that given by Kay for the length of time which elapses before gumbotil begins to form.

Kay subtracts 50,000 years from his estimates of the length of the Sangamon, Yarmouth, and Aftonian interglacial stages, assuming that that much time must elapse before gumbotil begins to form, and then, dividing the remaining times by the thickness of the gumbotils upon the Illinoian, Kansan, and Nebraskan drifts, reaches the conclusion that 8 feet of Nebraskan gumbotil formed in 150,000 years, 12 feet of Kansan gumbotil in 200,000 years, and 5 feet of Illinoian gumbotil in 70,000 years. These estimates give values ranging from 1 foot in 18,000 years to 1 foot in 14,000 years.

An estimate of the rate of formation of gumbotil in Indiana may be made by comparing the thickness of Horizon 2 (the horizon in which gumbotil and mesotil form) in the buried Illinoian drift with the thickness of that horizon in the area in Indiana where the Illinoian drift was not covered by Wisconsin drift, loess, or lake silts. Where not covered, the Illinoian drift has been subjected to the weathering processes since the withdrawal of the ice sheet. We would expect Horizon 2 to be thicker in it than in the buried Illinoian drift, which was exposed to weathering only throughout Sangamon time. As stated above, about $4\frac{1}{2}$ feet of gumbotil developed in Indiana during Sangamon time. From the study of numerous soil profiles of the Illinoian drift where not covered by Wisconsin drift,

loess, or lake silts, an estimate of 7 feet is made for the thickness of Horizon 2. This would mean that this horizon was thickened by 7 minus $4\frac{1}{2}$ feet, or $2\frac{1}{2}$ feet, during and after the deposition of the Tazewell drift and Peorian loess. If this time is taken at 45,000 years, then the additional $2\frac{1}{2}$ feet of gumbotil or mesotil was formed at the rate of 1 foot in 18,000 years.

While 45,000 years is probably a fair estimate for post-Tazewell time, along the lower Wabash Valley the Illinoian drift was probably covered somewhat earlier. The leaching of the loess there points toward the conclusion that loess accumulation began there as early as the Iowan substage, although the bulk of the loess may have deposited somewhat later. If loess accumulation began 50,000 years ago, then we have a rate for the formation of gumbotil of 1 foot in 20,000 years.

If we use the rate of 1 foot of gumbotil in 19,000 years, the average of the two estimates, we arrive at the following estimate of the length of Sangamon time. We assume that about 50,000 years elapse before gumbotil first appears. It would then require 85,000 years for $4\frac{1}{2}$ feet of gumbotil to develop, which would give a minimum estimate of 135,000 years for Sangamon time. This is 15,000 years greater than Kay's estimate but of about the same magnitude.

SUMMARY AND CONCLUSIONS

The Illinoian drift in southern Indiana has been leached to an average depth of 13 feet, and the Tazewell drift to a depth of $4\frac{1}{2}$ feet. This pronounced difference in the depth of leaching makes it possible to map accurately the boundary between the two drifts.

During Sangamon time the Illinoian drift was leached to an average depth of 6 feet, and 4–5 feet of gumbotil were formed. The thinner weathered zone in the buried Illinoian drift in Indiana, compared with that in Iowa and Illinois, suggests that the drift plain in Indiana during this time must have been one of very low relief, poor drainage, and high water table.

The silts of southwestern Indiana are of four types: local loess and lake silts of Illinoian age and extensive loess and lake silts of Wisconsin age. The bulk of the loess is equivalent to the Peorian loess in the states to the west. Correlatives of the Tazewell and

Iowan loesses of Illinois occur in Indiana within the area of the Tazewell drift. No late Sangamon loess such as has been described in Illinois has been recognized in Indiana. The Peorian loess is leached nearly $5\frac{1}{2}$ feet, which indicates that, while no Iowan drift is present in Indiana, loess accumulation probably began there as early as the Iowan substage.

An estimate of 45,000 years is made for the length of time that the Tazewell drift has been undergoing leaching. An estimate of 135,000 years is made for the duration of Sangamon time in southern Indiana. It is believed that these figures represent minimum, rather than maximum, estimates.

THE LOW AND BALL OF THE EASTERN SHORE OF LAKE MICHIGAN

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ABSTRACT

Lows and balls constitute a series of submerged troughs and ridges built approximately parallel to beaches with an abundant supply of sand and a gently sloping bottom. They are the work of plunging breakers and form best where the shoreline is relatively straight and where the prevailing winds are onshore. They remain in approximately the same position as long as conditions do not change. If change in wave size (height or length) occurs, the lows and balls may move slightly onshore or offshore as the position of the breaker line changes. However, if the water-level lowers, or other conditions change so as to bring the crest of the ball near the surface, currents and waves of translation are set up across the top of the ball, and the structure then migrates shoreward as a subaqueous dune.

The currents and waves of translation across the top of the ball prevent its being built up above the water surface by the work of the waves alone. Thus the ball is not the immediate forerunner of the offshore bar or barrier beach. On the inshore terrace, shoreward of the true lows and balls, ridges are built in various ways by waves and currents and frequently are driven to shore as subaqueous dunes. It is these minor ridges that have been mistaken by some investigators for migrating balls.

Lows and balls constitute a series of troughs and ridges at various places along the shores of the oceans and the larger lakes where the bottom slopes gently outward and where there exists an abundant supply of sandy sediment. These troughs and ridges lie roughly parallel to the shoreline and to each other and may extend continuously for many miles along the coast (Fig. 1). Such subaqueous troughs and ridges have been mentioned occasionally in geological literature for nearly three-quarters of a century and have been described from various parts of the world. Although they are of considerable interest because of their possible bearing on the origin of such beach structures as spits and offshore bars, their formation and behavior have been studied only along the shores of the Baltic Sea by a few German geologists and geographers.

In referring to certain subaqueous ridges along the British coast the English investigators, Wheeler,[1] Cornish,[2] and Gress-

[1] W. H. Wheeler, *The Sea-Coast* (London, 1902), pp. 41 and 118.

[2] Vaughn Cornish, "On Sea-Beaches and Sand-Banks," *Geog. Jour.*, Vol. XI (1898), p. 637.

well,[3] used the terms "low and ball" and "low and full"; the term "full" being used to designate the ridge formed by the waves between high and low tide and the term "ball," to designate the ridge formed below low-tide level.

Recently Kindle,[4] in describing a series of ridges on the bottom of Chesapeake Bay which lie parallel to the shore, used the term "ridge

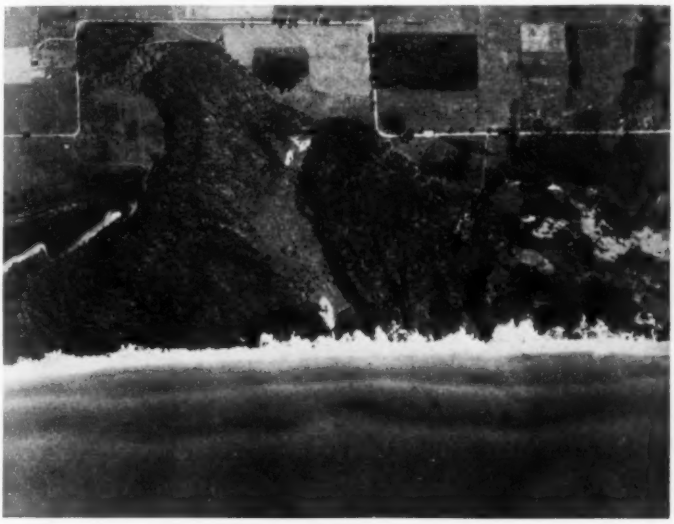

FIG. 1.—Aerial view of lows and balls

and trough" structure. He states that these structures are developed very extensively around the shores of Chesapeake Bay and that similar forms are found in Lake Ontario at Fair Haven, New York. The slope of Chesapeake Bay is very gentle. At Chesapeake Beach it is described as 10–15 feet per mile for 3 miles from the eastern and western shores. In one place he says the troughs are spaced "at in-

[3] R. Ray Gresswell, "The Geomorphology of the Southwest Lancashire Coast-Line," *Geog. Jour.*, Vol. XC (1937), p. 336.

[4] E. M. Kindle, "Notes on Shallow Water Structures," *Jour. Geol.*, Vol. XLIV (1936), pp. 863–67.

tervals of 85 feet, the axes of the ridges rising 2 and 3 feet above the troughs."

Kellogg[5] speaks of subaqueous ridges off the coast of Louisiana and states that he waded 1,500 feet across a series of low troughs and ridges with hard bottom. He says: "The crests of these ridges were from 10 to 20 feet across and they were from 20 to 30 yards apart." From this it appears that here as on the Chesapeake the water is shallow and the ridges comparatively small and numerous.

The subaqueous ridges along the shores of Lake Michigan are larger and farther apart than those mentioned by Kindle and Kellogg. On clear days they can be seen extending for long distances parallel to the shoreline. The attention of several American geologists has been attracted to them. They were mentioned by Desor[6] as early as 1851, by Andrews[7] in 1870, and again by Whittlesey[8] in 1886. Later the lows and balls of Lake Michigan were described by I. C. Russell[9] in some detail, and they were also noted by Gilbert,[10] who made some suggestions as to their origin.

Probably the earliest description of low and ball was given by Hagen in his *Handbuch der Wasserbaukunst*, which was published in 1863. He explained the formation of the ridges by shoreward-moving waves meeting the undertow, especially if the undertow is reinforced by backward-moving water in the troughs of oscillation. The lows and balls along the shores of the Baltic Sea were later studied by several German geologists and geographers. Among the

[5] James R. Kellogg, "Notes on Marine Mollusks of Louisiana," *Gulf Biol. Sta. Bull. No. 3* (1905), p. 38.

[6] E. Desor, in J. W. Foster and J. D. Whitney, *Report of the Lake Superior Land District*, Part II (Washington, D.C.: U.S. Gen. Land Office, 1866), p. 258.

[7] E. Andrews, "The North American Lakes Considered as Chronometers of Postglacial Time," *Trans. Chicago Acad. Sci.*, No. 2 (1870), p. 14.

[8] Charles Whittlesey, "Freshwater Glacial Drift of the Northwestern States," *Smithsonian Contrib. 197* (1896), p. 17.

[9] I. C. Russell, "Geological History of Lake Lahontan," *U.S. Geol. Surv. Mono. 11* (1885), pp. 92–93.

[10] G. K. Gilbert, "The Topographic Features of Lake Shores," *U.S. Geol. Surv. 5th Ann. Rep.* (1885), p. 111; "Lake Bonneville," *U.S. Geol. Surv. Mono. 1* (1890), p. 44.

better known of these investigators are Lehmann,[11] Braun,[12] Otto,[13] Passarge,[14] and Hartnack.[15] All of them except Otto believe the lows and balls to be quite unstable structures and subject to frequent changes in position. They also think that frequently the ball nearest the shore is driven in to become part of the beach ridge, or that it may sometimes be built above the water before reaching shore.

Otto, Passarge, and Hartnack all recognized that the lows and balls are the work of the plunging breakers, although none of them clearly comprehended the mechanics of the process. Passarge thought that in forming the low most of the material is thrown toward the shore rather than away from it. On the other hand, Otto gave as his opinion that the trough is not on the sea side of the ridge as indicated by Passarge but on the land side of it.

All the German investigators, with the exception of Otto, seem to have been unfortunate in choosing the vicinity of harbors for their investigations. The aerial photographs of the eastern shore of Lake Michigan show clearly that where piers or breakwaters project a few hundred feet into the lake the lows and balls are always more or less disturbed and broken up for a half-mile in each direction along the shore; thus, such an area furnishes a very poor field for study. Otto, however, was more fortunate in his selection of a location. His study, which extended over a period of 5 years, was made along the coast in the vicinity of Prerow. By means of soundings he took a series of fifty-two profiles perpendicular to the shoreline. These were taken every 250 meters for a distance of 12.75 kilometers.

Although Otto discovered many important facts about lows and balls, he failed to give any satisfactory explanation of their origin.

[11] F. W. P. Lehmann, "Das Küstengebiet Hinterpommerns," *Zeits. d. Gesell. f. Erdkunde zu Berlin*, Band XIX (1884), p. 391.

[12] G. Braun, "Entwickelungsgeschlichtliche Studien an Europäischen Flaschlandsküsten und ihren Dünen," *Veroff. Inst. f. Meerskunde, etc.*, Band XV (1911), pp. 89–91.

[13] Theodor Otto, "Der Darss und Zingst," *Jahrb. d. Geog. Gesell. zu Greifswald*, Band XIII (1911–12), pp. 393–403.

[14] Seigfred Passarge, "Physiologische Morphologie," *Mitt. d. Geog. Gesell. in Hamburg* (1912), pp. 195–98.

[15] Wilhelm Hartnack, "Über Sandriffe," *Jahrb. d. Geog. Gesell. zu Greifswald*, Band XL/XLII (1924), pp. 47–70.

He believed new balls are started by extraordinarily strong storms and that most of the changes in the lows and balls take place during the very violent storms. He thought the ball nearest the shore is formed where the incoming waves meet the backwash but offered no explanation of the origin of the outer balls, although he believed that there is a direct relation between storm-wave length and distance between the balls.

Johnson[16] includes the American, English, and German structures all together under the head of "Low and Ball," although in the last paragraph of his discussion he makes the following statement:

> In classifying the forms observed in the Great Lakes by Desor, Gilbert, Russell, and others, with those observed on tidal shores by Cornish and Wheeler, and giving the English name of low and ball to the series, I have proceeded on the assumption that the two forms are similar in character and identical in origin, such differences as noted being due to the changing water level in the case of the marine type. I must state, however, that this procedure is not based on any careful comparison of the forms as developed in the lakes and on the ocean, and my classification is accordingly to be accepted with due reservation.

There seems little room for doubt that the structures studied by the Germans along the shores of the Baltic Sea and commonly designated by them as *Sandriffe* are the same as those found along the east shore of Lake Michigan. From the descriptions given by Wheeler and Cornish it also seems probable that the ridges described by them are similar in origin to those of Lake Michigan and the Baltic except that the greater rise and fall of the tide on the English coast may modify the process to some extent. Therefore the use of the expression "low and ball" as a generic term to designate such structures seems justified.

THE REGION STUDIED

Lake Michigan from the city of New Buffalo to Grand Traverse Bay was selected for study because along this coast the low and ball structures are found nearly everywhere and are exceptionally well developed. The shore has the form of an extended S-shaped curve with a north and south trend. As far north as Little Point Sable it is quite regular in outline, and the bottom of the lake slopes more

[16] D. W. Johnson, *Shore Processes and Shoreline Development* (New York: John Wiley & Sons, Inc., 1919), pp. 486–89.

gently out from shore than it usually does farther north. From Little Point Sable to Traverse Bay the coast is more irregular, although its curves are mostly uniform and gentle. Sand is very abundant everywhere along the beach.

Throughout the whole distance the material of the coast is glacial drift which varies in composition from clay to coarse gravel and boulders. Even where the material is clay there is considerable sand on the beach. The greater part of this sand is the result of the waves cutting into the loose material of the shore. Very little comes from entering streams since most of the stream mouths are drowned. Considerable areas of sand dunes have formed in many places, especially on low shores near the drainage outlets.

On a bright day the balls can often be seen as light-colored streaks parallel with the shore. They also show very clearly in the photographs (Fig. 1). The average number of balls shown in the pictures is three. There are rarely more than four, and in a few places where the water deepens rapidly only one is present.

In investigating the lows and balls five methods of attack were employed.

1. *Study of aerial photographs.*—A study of the lows and balls was made from aerial photographs and for this purpose all the Agricultural Adjustment Administration offices from Berrien County to Leelaunau County were visited. This study was later supplemented by a brief study of aerial photographs on file at Lansing, Michigan, in the state Agricultural Adjustment Administration office and in the office of the State Geological Survey. The photographs gave much information regarding the relations between the lines of lows and balls and between them and the shoreline. As much ground can be covered in this way in a few days as in many summers' work with the sounding line.

The number of lows and balls shown in the photographs varies from one to four. In no case are more than four ridges visible in a picture, but this does not prove that others may not be present and concealed by greater depth of water. As stated, a series containing a greater number was seen by Kindle in Chesapeake Bay and by Kellogg along the Louisiana coast. In both these cases, however, the water is shallower and the bottom has a more gentle slope.

The spacing of the balls is somewhat uneven, and in general the distance between the second and third is greater than that between the first and second (see profiles, Fig. 2). The ridges are more regular in number and spacing along the relatively straight, uniform shores to the south of Point Betsie than to the north where the depth of water is more variable and the shoreline changes direction more frequently. The edges of the balls are not always clearly defined in the pictures. Sometimes the shoreward edges appear sharper than the outer edges. In some places there are long projections running out at an angle from the sides of the balls similar to the subaqueous ridges which form a part of the beach cusps of the second class.[17] These are particularly noticeable in West Traverse Bay where the currents set up by the waves are very strong and irregular. Similar subaqueous structures have been noticed in some places along the Atlantic coast.[18]

The outer balls usually extend for long distances without having any connection with the shore. Where they do show any dependence on the shore it is through becoming united with the No. 1 ball or by displacing it. The second ball was traced on the aerial photographs from the Pentwater piers to the Whitehall piers. Throughout this distance of 30 miles it maintained a fairly uniform distance from the shore and did not touch the ball either in front or back of it. This indicates that the material in the balls is not brought directly from the shore by littoral currents. It also shows that little or no shoreward migration of the balls occurs because a migrating ridge of that length lying offshore and passing out around Little Point Sable would not hold an almost constant distance from the shore. It would surely move faster in some places than in others because of inequalities in winds, waves, and currents and so would be united with the shore in some places and would be relatively far away in others. Russell evidently noted this characteristic of the balls for he says: "They can be traced continuously for hundreds of miles."

The following are a number of local peculiarities of the lows and balls discovered by studying the aerial photographs.

[17] Evans, "The Classification and Origin of Beach Cusps," *Jour. Geol.*, Vol. XLVI (1938), pp. 618–19.

[18] Major A. C. Lieber, Jr., member of Beach Erosion Board, private communication.

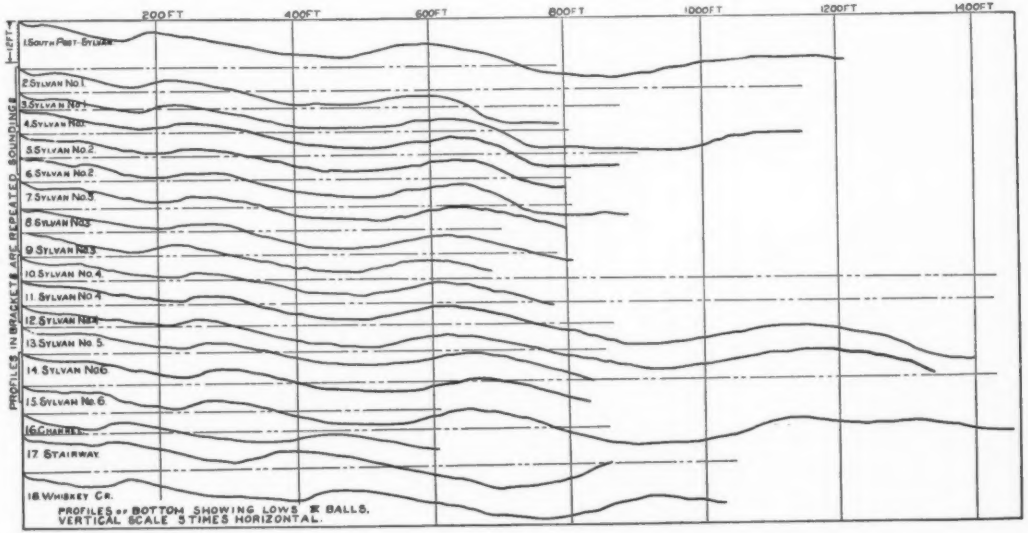
200FT
400FT
600FT
800FT
1000FT
1200FT
1400FT
PROFILES IN BRACKETS ARE REPEATED SOUNDINGS
1 LEFT
1. South Fast-Sylvan
2. Sylvan No 1.
3. Sylvan No 1
4. Sylvan No.
5. Sylvan No 2.
6. Sylvan No 2
7. Sylvan No 3
8. Sylvan No 3
9. Sylvan No 3
10. Sylvan No 4
11. Sylvan No 4
12. Sylvan No 4
13. Sylvan No 5
14. Sylvan No 6
15. Sylvan No 6
16. Channel
17. Stairway
18. Whiskey Cr.
PROFILES OF BOTTOM SHOWING LOWS & BALLS.
VERTICAL SCALE 3 TIMES HORIZONTAL.

FIG. 2

The inner ball may join the shore. Where it does it is always at an acute angle. Ball No. 2 thus becomes ball No. 1. Nowhere were two balls seen to join the shore at the same place (Fig. 3, *A*).

In some places two balls join and become one. Where that happens another ball usually comes in to maintain the number (Fig. 3, *B*).

Where there is a rounded projection of the shoreline the balls usually parallel the shore and each other as shown in Figure 3, *C*. A similar arrangement was observed at the old entrance to Saugatuck Harbor. This entrance has been closed for many years, and the capelike projection caused by accretion around the old piers has caused the balls to move outward (Fig. 3, *D*). It will be noted in this case as well as the one mentioned above that the lows and balls follow around the projecting shoreline and neither join it nor are thrown offshore as they probably would be if they were built by littoral currents.

Usually where a ball joins the shore it makes a very acute angle with it, but about a mile north of Pentwater piers ball No. 1 appears to turn and join the shore almost at a right angle (Fig. 3, *E*). Again at the outlet of Bass Lake the balls seem to break squarely across (Fig. 3, *F*). This breaking-across is probably caused here by the extra supply of sand coming from Bass Lake outlet.

In some places two balls are joined for a short distance. This occurred about a mile south of Bass Lake outlet (Fig. 3, *G*).

A quarter-mile south of Silver Lake outlet a temporary ridge several hundred feet long was found to unite ball No. 1 with the shore (Fig. 3, *H*).

2. *Measurements from the shore.*—The method of sighting on the balls from the shore was very simple but it proved to be effective and sufficiently accurate for the purpose. This was to determine whether or not the balls moved with respect to a fixed point on the shore and, if so, in which direction.

A number of easily recorded points were selected along the shore. As far as possible preference was given to points having a considerable elevation above the lake. The observations were made by standing on a selected point and holding an ordinary carpenter's rule vertically at arm's length. The top of the rule was held even with the

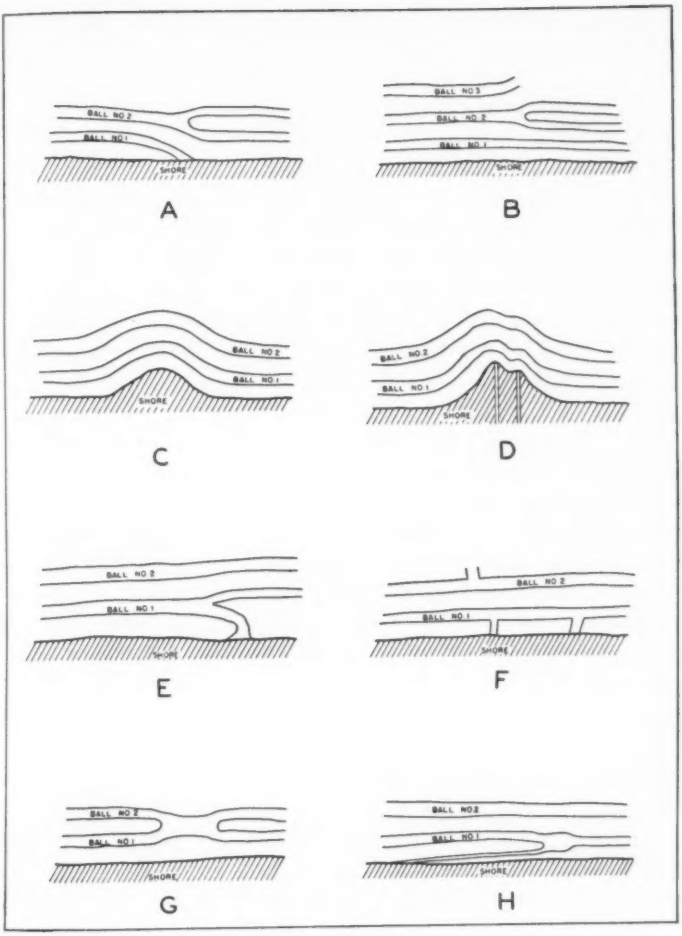

FIG. 3.—Features of balls

line of sight to the water horizon, and the end of the thumb was moved up or down on the scale until it was in the line of sight to the ball. The distance on the scale between the end of the thumb and the top of the rule was recorded. By repeating the observations over a considerable period of time it was possible to determine the movement, if any, of the position of the balls.

The observations were begun in June of 1938 and continued through the summer of 1939. They were taken at various places along the shore from Muskegon north to the High Stairway, a distance of about 18 miles. The results are given in Table 1. Ball No. 1 is the one nearest the shore at the time of observation. Nowhere were more than three balls visible during the observations. Sometimes conditions were such that No. 3 was not visible and occasionally only No. 1 could be seen.

A variation of $\frac{1}{16}$–$\frac{1}{8}$ inch will be noticed in some of the readings. Small variations are to be expected with the method used because of refraction. When the air is warmer than the water the horizon will appear too high, thus causing the readings to be slightly too large.

The actual elevations of the points of sight above the lake are not given since they are of little significance in the measurements. The figures are relative and have meaning only through comparing successive readings. An increase in the size of a number indicates movement of the ridge toward shore. The direction of the line of sight was always controlled by sighting over a fixed point situated between the point of observation and the lake.

Only one series of observations was recorded for 1938. Other sights were taken in the latter part of August of that year, but no record was made of them except to note that no change occurred in the position of the balls. At that time it was thought that the winter storms would cause a decided change in the low and ball positions and this was expected to show in the 1939 observations. A question mark in the table indicates that a reading was not obtainable on that date because of poor visibility.

At Duck Lake bridge on July 28, 1938, the position of ball No. 1 is indicated by a reading of $\frac{1}{2}$ inch on the scale. This reading is the same as that for ball No. 2 on June 28, July 30, and August 15, 1939. However, a new ridge is shown on the latter three dates by the

TABLE 1

Date	Duck Lake Bridge	Duck Lake Pavement	Duck Lake Tree	Old Channel	Gray's Landing	North Muskegon	High Stairway
7-28-38	1 = 1/2″ 2 = 5/16 3 =	1 = 7/8″ 2 = 7/16 3 =	1 = 2 1/8″ 2 = 1 1/4 3 =	1 = 2 9/16″ 2 = 1 9/16 3 =	1 = 5 5/8″ 2 = 3 7/8 3 =	1 = 1 9/16″ 2 = 3/4 3 =	.....
6-28-39	1 = 7/8″ 2 = 1/2 3 =	.....	1 = 3 1/4″ 2 = 2 1/8 3 =	.....	.....	.....	First sight
7-9-39	.....	.....	.....	1 = 3 9/16″ 2 = 1 9/16 3 = 7/8	1 = 5 3/4″ 2 = 3 1/2 3 = 1 5/16	.....	.....
7-15-39	.....	.....	.....	1 = 3 1/4″ 2 = 1 9/16 3 =	.....	1 = 1 3/4″ 2 = 3 =	.....
7-30-39	1 = 1 3/16″ 2 = 1/2 3 =	1 = 1 15/16″ 2 = 31/32 3 = 1/2	1 = 3 5/16″ 2 = 2 9/16 3 = 1 5/16	.....	.....	.....	.....
8-12-39	.....	1 = 1 7/8″ 2 = 7/8 3 =	1 = 3 5/16″ 2 = 2 9/16 3 =	.....	.....	.....	.....
8-15-39	1 = 1 3/16″ 2 = 1/2 3 = 5/16	1 = 1 15/16″ 2 = 1 3 = 1/2	1 = 3 1/2″ 2 = 2 9/16 3 = 1 5/16	1 = 3 1/16″ 2 = 2 7/16 3 = 1 9/16	.....	1 = 1 11/16″ 2 = 1 1/16 3 =	No change
8-17-39	.....	.....	.....	1 = 2 11/16″ 2 = 1 9/16 3 = ?	.....	.....	.....
9-1-39	.....	.....	1 = 4 1/8″ 2 = 2 1/4 3 = 1 5/8	1 = 3 5/16″ 2 = 1 9/16 3 = 3/4	1 = 5 5/8″ 2 = 3 5/16 3 = 1 1/16	.....	.....
9-9-39	.....	.....	1 = 2 5/16″ 2 = 1 3/8 3 =	.....	.....	1 = 1 11/16″ 2 = 3 =	.....
9-14-39	.....	.....	1 = 2 3/16″ 2 = 1 9/16 3 =	1 = 3 1/16″ 2 = 1 5/8 3 =	1 = 5 1/4″ 2 = 3 3/8 3 = 1 3/4	.....	No change
10-21-39	.....	.....	1 = 2 3/32″ 2 = 3 =	.....	.....	1 = 1 11/16″ 2 = 3 =	.....

figures $\frac{7}{8}$ inch, $1\frac{3}{16}$ inch, and $1\frac{9}{16}$ inch. There is a possibility that the ball seen as No. 1 in 1939 might have been overlooked in 1938 but this is not probable because it was not seen when later unrecorded observations were made in 1938. Also ball No. 2 in 1938 gave a reading of $1\frac{5}{16}$ inch, which is the same as recorded for ball No. 3 in 1939.

The Duck Lake pavement observation point is a few hundred feet south of the Duck Lake bridge, and the results of the readings here substantiate those obtained from the bridge.

The Duck Lake tree is about $\frac{1}{4}$ mile south of the Duck Lake bridge. Here again the positions of the balls Nos. 1 and 2 on July 28, 1938, are the same as that of balls Nos. 2 and 3 at the time of the first four observations in 1939. The same condition also probably held for September 1, 1939, since the sight on that day was taken late in the afternoon, which had a tendency to make the readings larger because of refraction.

On August 9 there was a heavy gale which was followed by several days of stormy weather. During this time the inner ridge of material at Duck Lake moved in toward shore and by September 9 it was so close in as to be no longer visible from the points of observation. This was the temporary ridge of June 28 to September 1 moving toward shore while the true balls remained fixed in position.

The station at the stairway north of Old Channel is more than 2 miles north of Duck Lake. The readings here show a little more extensive series of changes in the sediments near shore than was seen at Duck Lake, but here again the well-established balls remained in position while the inshore changes occurred.

At Gray's Landing all lows and balls held a relatively constant position. No. 1 may have moved out a little and then come in again while No. 2 moved out slightly from July 28, 1938, to August 4, 1939.

At North Muskegon the outer balls were seldom visible because of turbidity of the water. However, the figures obtained are in good agreement. Ball No. 1 appeared to be a little closer inshore in 1939 than in 1938.

At the High Stairway about 13 miles north of White Lake piers the only sights taken were in 1939. These were obtained from an observation tower high above the lake. The sights were taken by

placing a yardstick against the side of the tower and sighting along its edge to the balls.

The foregoing figures indicate that for the period of about 14 months covered by the observations little or no movement of the well-established balls occurred. It also shows that close to shore in some places ball-like structures did form and move rather freely, this movement being generally shoreward.

3. *Soundings and studies of profiles.*—Lines of soundings running from the shore outward across the lows and balls were made at various selected localities from about a mile south of White Lake piers to about 12 miles north of them. Near the south end of Sylvan Beach, about a mile south of White Lake piers, a sounding field was laid out and repeated soundings were taken during the months of July and August, 1939.

The method of work was as follows: A $\frac{1}{4}$-inch manila line was marked every 10 feet. One end of the line was made fast to a stake on the shore at a known distance from the water's edge. The other end was then carried out perpendicular to the shoreline and made fast to a buoy. The main measuring line used was 1,000 feet long, as this was found to be the greatest length that could be handled effectively from a rowboat. A smaller graduated line and a second buoy were used for taking soundings farther out. The sounding line was marked at 1-foot intervals and the tenths were estimated as the soundings were taken. The lead is not hauled up between soundings but is merely lifted free of the bottom and carried to the next point. By running the boat slowly it is possible to take soundings in this way very accurately. Smooth, quiet water is necessary because soundings of the required accuracy cannot be taken from a pitching boat, and if much current is running it is impossible to prevent drifting.

The repeated soundings taken near Sylvan Beach were for the purpose of determining what changes, if any, occur on the lows and balls during a short period of time. In the beginning of the work it was expected that such changes might be numerous and of considerable magnitude. It was soon found, however, that the changes are small in amount and not very rapid. For this reason as well as because of rough, unfavorable weather only three successive sets of soundings were taken. The results of the work are shown in Table 2.

TABLE 2

	Ball No. 1						Ball No. 2					Ball No. 3				
	Shore Zone Slope	Trough Depth	Crest Depth	On-Shore Slope	Off-Shore Slope	Dis. From Shore	Trough Depth	Crest Depth	On-Shore Slope	Off-Shore Slope	Dis. From Shore	Trough Depth	Crest Depth	On-Shore Slope	Off-Shore Slope	Dis. From Shore
1. South Post, Sylvan Field (7–27)	1/22	6.0	3.4	1/13	1/30	200	11.0	7.3	1/24	1/18	600	18.0	12.0	1/33	.....	1,120
2. Sylvan No. 1 (7–17)	1/22	5.8	3.4	1/18	1/24	215	11.0	8.3	1/30	1/11	620	.....	.....	.....	.....	.....
3. Sylvan No. 1 (7–27)	1/25	6.0	4.0	1/20	1/24	225	11.1	8.4	1/35	1/11	625	18.0	13.0	.....	.....	1,130
4. Sylvan No. 1 (8–16)	1/27	5.7	4.0	1/25	1/35	245	12.0	8.7	1/36	1/12	650	17.5	.....	.....	.....	.....
5. Sylvan No. 2 (7–18)	1/25	6.1	4.6	1/20	1/26	255	11.5	8.5	1/30	1/14	640	.....	.....	.....	.....	.....
6. Sylvan No. 2 (8–16)	1/17	6.0	4.8	1/20	1/23	260	12.0	8.5	1/27	1/12	640	18.0	.....	.....	.....	.....
7. Sylvan No. 3 (7–18)	1/24	6.2	4.8	1/25	1/22	265	11.8	8.1	1/25	1/25	645	.....	.....	.....	.....	.....
8. Sylvan No. 3 (8–16)	1/27	6.2	5.0	1/25	1/19	265	12.0	8.4	1/26	1/21	620	.....	.....	.....	.....	.....
9. Sylvan No. 3 (8–26)	1/24	6.0	3.8	1/20	1/23	230	12.0	8.8	1/20	1/30	600	.....	.....	.....	.....	.....
10. Sylvan No. 4 (7–18)	1/25	6.2	5.0	1/15	1/25	260	12.0	8.1	1/27	1/21	615	.....	.....	.....	.....	.....
11. Sylvan No. 4 (8–16)	1/27	6.0	5.0	1/10	1/21	275	12.0	8.0	1/27	1/24	625	18.4	13.0	1/30	1/29	1,160
12. Sylvan No. 4 (8–26)	1/35	5.9	4.0	1/22	1/25	260	12.0	8.5	1/21	1/24	630	19.0	13.1	1/30	1/23	1,175
13. Sylvan No. 5 (7–19)	1/25	6.1	5.0	1/15	1/23	285	11.4	8.0	1/33	1/25	660	.....	.....	.....	.....	.....
14. Sylvan No. 6 (7–19)	1/24	7.0	4.9	1/15	1/23	290	11.5	8.2	1/25	1/19	680	.....	.....	.....	.....	.....
15. Sylvan No. 6 (8–18)	1/20	7.0	4.8	1/20	1/21	290	12.0	8.2	1/22	1/17	660	19.7	12.0	1/19	.....	1,250
16. Old Channel (7–27)	1/25	4.9	3.1	1/18	1/28	145	9.2	7.0	1/20	1/29	450	.....	.....	.....	.....	.....
17. Stairway (7–19)	.....	3.0	2.3	1/25	1/23	140	9.0	6.0	1/20	1/23	395	17.9	11.0	1/17	.....	890
18. Whiskey Cr. (8–5)	1/25	4.0	2.0	1/11	1/23	155	9.7	6.8	1/15	1/25	480	16.0	10.0	1/21	.....	935
Maximum	1/17	7.0	5.0	1/10	1/19	290	12.0	8.8	1/15	1/11	680	19.7	13.1	1/17	1/23	1,175
Minimum	1/35	3.0	2.0	1/25	1/30	145	9.0	6.0	1/36	1/30	395	16.0	10.0	1/30	1/29	890
Average	1/25	5.8	4.1	1/19	1/24	237	11.3	8.0	1/26	1/21	602	18.1	12.0	1/24	.....	1,094

The lines of soundings varied in length from 610 to 1,460 feet. The shortness of some of the lines is due to winds rising at the time the work was being done, which made it impossible to continue.

Each ball is accompanied by a low *shoreward* of it. The ball with its low in front constitutes a unit. This relation between the low and ball is contrary to the belief of Cornish, who says: "The building up of a Full of sand in front of the breaker is accompanied by the excavation of a trough, or Low, at the back of the breaker.[19] He was probably led into this error by his belief that the low and ball constitutes a giant ripple, since he states: "This is roughly similar to the simultaneous excavation and elevation which produces the ridge and furrow so well known as 'ripple-mark.' " Wheeler seemed to consider the low as being shoreward of the ball, as he says: "The sandy beaches frequently consist of wide ridges and lows, the lowest places generally being found where the waves plunge at high water spring tide."[20] As before indicated, the German investigators also failed to grasp the true relation of the lows and balls to each other.

Table 2 shows that the average distance between the balls increases with distance from the shore. The increase in depth of the bottom with each succeeding low is about 6 feet while that of the crests of the balls is 4 feet. The slopes of the sides of the balls is much the same for all the ridges, the average variation in slope being from 1 in 19 to 1 in 26. The maximum slope obtained, 1 in 10, was found on the shoreward slope of ball No. 1. This is 5.7°, which is much less than the subaqueous angle of repose for sand of the Lake Michigan beach. The least slope is 1 in 36 or 1.6°. Evidently the size of the balls increases with distance from the shore.

Fourteen of the 18 profiles of ball No. 1 are steepest on the shoreward side. Over ball No. 3 there are hardly enough measurements to warrant a comparison, but those available show no great uniformity of slope in either direction.

Otto gives the results of four measurements over balls Nos. 1 and 2, and two over the third ball. Out of the 10 sections he found 7 in which the slope was steeper on the land side. He concludes as follows:

> Profile No. 1 fills throughout the condition of a steeper landside and a flatter seaside. Profile No. 2, on the contrary, overthrows this entirely. It shows exact-

[19] *Op. cit.*, p. 637. [20] *Op. cit.*, p. 41.

ly inverted proportions and disproves entirely that there is any fixed rule regarding the proportions. Profiles 3 and 4 substantiate this. However one may say that in the majority of cases the landside is somewhat steeper than the seaside.[21]

Hartnack,[22] who studied three series of profiles taken near Swinemünde in 1898, 1903, and 1908, is also of the opinion that on the average the balls have their steeper sides toward the land.

I. C. Russell, in his description of lows and balls on Lake Michigan, says:

> There are usually two, but occasionally three, distinct sand ridges; the first being about 200 feet from the land, the second 75 or 100 feet beyond the first, and the third, when present, about as far from the second as the second is from the first. Soundings on these ridges show that the first has about 8 feet of water over it, and the second usually about 12; between the depth is 10 to 14 feet.[23]

It is quite probable that Russell's observations were taken near Sleeping Bear where the water deepens more rapidly than at Sylvan Beach. This may account for the greater depths over the balls and the lesser intervals between them.

It seems probable that the shallower the water and the more gradual the slope of the bottom the smaller are the lows and balls and the nearer they come to the surface. For example, the lines of soundings numbered 16, 17, and 18 were in shallower water than those taken at Sylvan Beach. This effect of the shallowing is also indicated by Kindle's[24] work on Chesapeake Bay where, in water that reached a depth of 7 feet or less at 1,000 feet from shore, there are sometimes as many as 10 balls spaced about 85 feet apart. It will be recalled that Kellogg found a somewhat similar condition off the coast of Louisiana.

An examination of Table 2 and of the profiles in Figure 2 shows that the shoreward slopes of ball No. 1 are somewhat steeper than are corresponding slopes on balls Nos. 2 and 3. This may be accidental or it may be, as perhaps in profiles Nos. 16, 17, and 18, that the ridges are not completely fixed. If such is the case the crest of the balls may have been driven slightly forward by the upper part of the breakers, as explained elsewhere, and the shoreward side of the

[21] *Op. cit.*, p. 395.
[22] *Op. cit.*
[23] *Op. cit.*
[24] *Op. cit.*

ridges steepened. Certainly the onshore and offshore slopes of the balls are not such as to indicate any general shoreward migration similar to the movement of unsymmetrical ripples.

It is evident in comparing the profiles resulting from successive soundings over the same course that slight changes occur in the shape of the lows and balls and in the distances of the ridges from shore. However, the exact crest is somewhat difficult to locate, and a change in the shape of the crests may give an apparent change of distance from shore.

On August 8, 1939, there was a heavy storm that lasted about 36 hours. The seas broke continuously for many hours over all three balls, and the sediments were stirred up to such an extent that the water was discolored some distance out beyond the third ball. The wind velocity during the storm, as estimated by the officer in charge of the Coast Guard Station at White Lake Harbor, was 35 miles per hour. During this storm, and the days of somewhat rough weather following, most of the changes in form and distance from the shore shown on the successive profiles took place (Fig. 2). However, no general migration of the lows and balls in either direction occurred. Soundings Nos. 3 and 4 show a movement away from shore of about 20 feet for both balls. There is no change of consequence between profiles Nos. 5 and 6. Profiles Nos. 7, 8, and 9 show a slight migration toward shore. Numbers 10, 11, and 12 may be said to be indeterminate. In Nos. 14 and 15 both ball No. 1 and ball No. 2 seem to have moved slightly shoreward. It seems from this that slight movements, sometimes shoreward and sometimes seaward, do occur. Whatever movements take place are probably most frequent in winter when storms are more severe, provided the lake is not frozen. This is also the opinion of the commercial fishermen who sometimes set their nets in the lows.

There are several possible causes that may bring about these changes. Currents over the balls are set up during times of strong winds and such currents sometimes have a high velocity. Any commercial fisherman of much experience has had his net buoys pulled under and his nets shifted by them. During a 15-mile wind which blew onshore at an angle of 45° a current of 4 miles per hour was measured in the inner low at Duck Lake beach. Winds of 40–60 miles

per hour must produce very strong currents, some of which may run across the balls in the lower places either toward or away from shore. Such currents were observed running across ridges in the inshore zone with a moderate onshore wind blowing.

A second cause of readjustment of the position of the lows and balls may be the variation of the water-level brought about by winds and differences in barometric pressure. As will be shown later, ridges built during times of high water may be driven inshore by moderate-sized waves when the water-level is again lowered.

A third process that may bring about a change of position of the balls as well as a change in their shape is a shifting of the position of the plunging breakers. When the average position of the plunging breakers is more on the offshore side of the ball it will cause the ball to move somewhat away from shore, but if the average position of the plunging breakers is shifted nearer the shore it will cause an onshore movement of the ball.

An examination of the profiles shows several cases of shifting of ridges in the inshore zone. In profile No. 7 the shore terrace extends out a distance of 90 feet. In No. 8 the sand from the edge of the terrace has been driven nearly to shore in the form of a ridge. In No. 9 it has reached the shore and there is an even slope from the shoreline out to the bottom of the first trough.

4. *Wave-tank studies.*—During 1938–39 a wave tank was constructed and operated at the University of Oklahoma with money furnished by the Faculty Research Fund. Among the phenomena observed during the operation of the tank the effect of the breaking waves stood out very prominently.

The tank was 20 feet long, 4 feet wide, and 22 inches deep. Sand was placed in one end to form a miniature beach. Oscillation waves were formed at the other end of the tank by using a typical wave machine having the faces of the generating trough set at an angle of 30°.

With uniform waves $3\frac{1}{2}$ inches high a trough was formed in 4 inches of water with its bottom $5\frac{1}{2}$ inches beneath the water surface. The formation of this trough was accompanied by the building of a ridge immediately on the offshore side of it with a crest 2 inches below the surface. This ridge, which was 40 inches from shore,

formed within a few minutes after starting the machine and maintained its position and depth for 3 hours, the whole time of the run. On another run using a 2-inch wave a trough and ridge were formed 31 inches from shore with its crest $1\frac{1}{2}$ inches below the surface. The trough shoreward of it was $2-2\frac{1}{2}$ inches deep.

On the third run waves 3 inches high were used. These caused a ridge to form having its crest 2 inches below the surface and a trough $3\frac{1}{2}$ inches deep on the shoreward side. This was 41 inches from shore. The machine was then changed to generate waves $\frac{7}{8}$ inch high. These passed over the trough and ridge already formed without disturbing it and formed a smaller trough and ridge 11 inches from shore. This ridge had its crest $\frac{3}{4}$ inch below the surface with a trough $1\frac{1}{2}$ inches deep shoreward of it. These parallel ridges may correspond to the low and ball, the outer one being formed by the larger waves and the inner one by the smaller waves which pass over the deeper ridge without disturbing it.

That the waves set up by the generating machine were oscillation rather than translation waves is indicated by the fact that pure oscillation ripples were formed in both deep and shallow water except in those places where definite currents were observed.

The ridges and troughs formed by the uniform waves of the wave tank were comparatively narrow in relation to their vertical extent when compared with similar structures on the lake bottom. This may be partly because the size and specific gravity of the particles of sediment in the tank have a very different relation to the size of the tank waves than does the sediment on the lake bottom to the size of the waves of the lake.

However, there are two conditions present in the case of the lake waves, but missing in the tank, which have much to do with the formation of the relatively broad ridges and troughs of the lows and balls. Often two or more series of waves of different sizes occur at the same time. When the crests of waves of two sizes happen to coincide the result is a larger wave which will break farther out in deeper water. When the crests do not coincide the result is smaller waves that either pass over the ball without breaking or else break nearer the shore than do the large ones. Also a series of wind-generated waves of oscillation do not have a phase velocity but have a

group velocity instead, which is only one-half that of the phase velocity.[25] This is the reason it is usually not possible to follow with the eye a single water wave for any great distance. The wave will seem to disappear. What it does is to merge with the wave in front or with the following wave. As a result of this merging the waves of a series are not all the same height since the troughs and crests may, or may not, coincide. This causes a variation in the position of the breaking waves.

This group action of the waves shows up strongly when an attempt is made to follow with the eye the movement of the waves from the outer ball to the next inner one. Some gliding breakers over the outer ball may be followed as waves which become plunging breakers on the next inner ball, but others appear to die out and it is the wave in front or behind that plunges on the inner ball.

It seems probable that the position of the low and ball is fixed at the place where the majority of the breakers plunge. A wave which plunges before reaching the inner edge of the ball will tend to destroy the ridge, but this damage is repaired by following waves which plunge more to leeward. A permanent change in the plunge line results in a shifting of the position of the ball until it is in adjustment with the new condition.

5. *Studies of the inshore zone.*—From the shore out about halfway to the first well-established ball is a zone in which the sediments move very erratically as compared with their behavior in the zone of the fixed lows and balls. Its outer margin is approximately where the depth of the water is equal to that covering the crest of the first ball. In this zone the bottom is undergoing almost constant change in both stormy and fair weather. Sediment is added to or taken away from the shore and is moved over the bottom and along the shore in either direction as the currents and waves vary with the changes in direction and force of the wind.

At the immediate shoreline the sediments are principally under the control of the currents in the swash and backwash. Farther out they are moved by the littoral currents, by the travel of the ripples set up by the waves of oscillation and translation, and by various

[25] C. G. Darwin, *The New Conception of Matter* (New York: Macmillan Co., 1931), p. 49.

other water movements, some of which are set up by waves and some of which are eddies due to surface movements of the water. In this zone there are frequent mass movements of sediments in which the behavior of the sand mass is somewhat similar to that of the sand in dunes.

When waves break on shallow submerged sand masses, translation waves and currents are set up across them and material is carried across in both suspension and traction. Also, unsymmetrical ripples are formed which travel slowly across the surface of the mass and carry material, just as material is transported in ripples along a stream bed. The material in the lee slope of the subaqueous dune stands at the angle of repose similar to that on the lee side of a sand dune. The subaqueous dune, however, has a much broader and flatter crest than does the sand dune and its altitude is probably never more than a few feet. As explained elsewhere, the leeward movement of such a sand mass seems to be determined by the depth of water over it and the ratio of this depth to the depth of water on each side of it. If the water is relatively deep over the sand mass, the waves may break without producing waves (currents) of translation of sufficient strength or depth to cause movement of the underlying sediments. In fact, if the water over the mass is relatively deep, plunging breakers will act on it and the mass will become relatively fixed in position. If the top of the mass is close to the water surface, however, the currents will cause the material to be carried rapidly to leeward.

These inshore sand masses which later may become moving subaqueous dunes have several different origins. During heavy storms the waves take material from the shore, but because of local differences of depth, shape, and hardness removal does not occur at the same rate everywhere along the shore. This results in a scalloped effect giving cusps of Class 1.[26] If the wind later blows obliquely onshore, beach drifting will be more rapid on the windward than on the leeward side of these cusps and any littoral current near shore will be deflected outward. As a result a ridge is built out under water from the extremity of a cusp and it becomes a cusp of Class 2, which has many of the characteristics of a spit. Sometimes these ridges may

[26] Evans, *op. cit.*, pp. 616–18.

extend a considerable distance parallel with the shore. Occasionally with a change of wind of about 90° such ridges are cut off from the shore and so become independent ridges of the inshore zone.

Material taken from the shore during strong gales is often deposited in the form of a shallow terrace just offshore (Fig. 2, profiles Nos. 7, 10, 13, and 14). Such terraces are particularly likely to form during times of rising lake level, such as occurred during 1938–39. Along the east shore of Lake Michigan these are usually 50–100 feet wide and have a depth at the outer edge of perhaps 2 or 3 feet. Waves of moderate height will sometimes break at the edge of such terraces and in so doing will build a ridge. This ridge has some of the characteristics of a permanent ball and with a rising water-level might become ball No. 1. However, in nearly all cases it is moved in toward the shore as a subaqueous dune. It is also quite probable that deposition occurs on the terrace where the backwash meets the incoming waves, thus causing them to drop sediment from suspension and so form a ridge. This process has been emphasized by many investigators since the water movements involved are easily seen from the shore.

Ridges formed in these various ways and later moved toward the shore have been mistaken for true balls in the process of being forced toward shore by the waves. This is probably what led many investigators to the mistaken conclusion that the balls migrate toward shore and reach it at more or less regular intervals.

A ball maintains its position because of the downward plunge of the breaking wave crests, the seaward movement of the water in the trough immediately in front of it, and the turbulence set up by the plunging wave. But if the crest of the ball be brought so near the surface that breaking waves form a current across the top, the sand mass becomes a subaqueous dune. Then the tops of the broken waves travel on as ridges of water with foaming fronts. Such waves have flat intercrest areas in which the water is either stationary or, in sufficiently shallow water, moving forward but at a less rate than the wave. This is what J. S. Russell[27] meant by oscillation waves being changed into waves of translation on approaching the shore.

Sometimes the currents crossing the sand mass are so strong that

[27] J. Scott Russell, "Report on Waves," *Rept. Brit. Assoc.*, Vol. XIV (1844–45), pp. 311–90, esp. p. 372.

the whole top of the ridge is in a condition of semisuspension. At such times much material is carried forward, no ripples are formed, and the top of the mass is smooth. With lower velocities current ripples form which transport sand to leeward. Thus under both conditions considerable material is always being carried shoreward.

A glance at the cross-sections of the lows and balls shows that the depth of water in the lows to leeward of the balls is in all cases something less than twice its depth on their tops. A leeward movement of the sand ridge apparently does not begin until the depth on its crest is about half that to leeward. It is frequently much less than this. The depth to leeward is accompanied by an equal or greater depth to windward of the ridge and is merely an indication that the top of the ridge is so high above the bottom that waves breaking on its windward side will cause waves and currents of translation across it. Thus the subaqueous dunes in Crystal Lake near Frankfort, Michigan, have a depth on top of 3–4 feet with 12 feet of water to leeward. In scores of measurements made on smaller moving sand masses whose lee slopes were at the angle of repose for sand under water the ratio was never less than 1 to $2\frac{1}{2}$. With ratios as great as this it is probable that the amount of water passing over the top of the subaqueous dune is insufficient to set up any appreciable induction eddy or reverse current to leeward.

All that is necessary to set a mass of sand moving in this way is that a strong current be set up across it and there be a chance for the water to get away to leeward. This generally means, in the case of the low and ball, that the water above the ball must in some way become so shallow that when the waves break, the top of the plunging crests will move forward as translation waves and currents and so bring about a forward travel of the sediments. There are several ways by which ridges may have their crests brought close enough to the surface for this migration to occur.

When the subaqueous ridge is attached to a cusp of Class 2, the shallowing of the ridge is progressive as sediment from the shore is carried out and deposited along it. In such cases the ridge will migrate toward shore at the same time it is building. This has been described by Evans.[28]

[28] "Mass Transportation of Sediments on Subaqueous Terraces," *Jour. Geol.*, Vol. XLVII (1939), pp. 327–28.

Sometimes at the end of a strong storm period when the water has been piled up high onshore the shoreward migration may begin at once as the fall in water-level brings the top of the ridge in reach of the upper part of the breakers.

Such changes in water-level sometimes occur on the Great Lakes. Hayford[29] shows that the combined effects of atmospheric pressure and onshore winds may raise the water-level at Buffalo and at Cleveland more than a foot. At Buffalo on October 21–22, 1910, the rise was 1.67 feet, and at Cleveland on October 22, 1910, it was 1.60 feet. It is quite probable that the east coast of Lake Michigan experiences rises in water-level of comparable size. Johnson[30] cites numerous instances in which the water along coasts has had its surface raised many feet by onshore winds and currents. During the hurricane of September, 1938, the water along the Rhode Island coast was elevated 10–15 feet above mean high tide by a combination of high spring tide and onshore wind.[31]

A glance at profiles Nos. 2 and 6 shows that the ridge at the edge of the terrace often has depth of about 2 feet at the crest and around 3 feet to leeward. If such a ridge were built up during high water, then the drop of 1 foot on the return to mean level would leave a depth of 1 foot at the crest and 2 feet to leeward. Under such conditions the incoming waves might cause the ridge to begin to migrate. Other ways in which sand ridges may be built up close to the water surface are explained on pages 497 and 498.

In order that the migration of subaqueous ridges on the terrace be carried to completion a nice balance of forces is necessary. If, during the time of the building and migration of the ridge in the inshore zone there is a rise in water-level and the waves increase in size and break strongly over the ridge, the sand will be thrown outward in the breakers and the ridge will temporarily become a stationary ball-like structure. It is during the time of lesser winds and comparatively

[29] J. F. Hayford, "Effects of Winds and Barometric Pressures on the Great Lakes," *Carnegie Inst. Pub. 317* (1922), Pls. 7–15.

[30] *Op. cit.*, pp. 125–26.

[31] R. L. Nichols and A. F. Marston, "Shoreline Changes in Rhode Island Produced by Hurricane of September 21, 1938," *Bull. Geol. Soc. Amer.*, Vol. L (1939), pp. 1357–90.

calm weather when the waves are relatively small that these ridges are moved toward shore and finally join the material of the beach ridge.

Thus it is, as has before been observed,[32] that during heavy storms the material of the beach is carried out and the shoreline moves landward, while during lesser storms and relatively calm weather material is brought back to the beach and the shoreline moved seaward. Hunt[33] states that this phenomenon has been indicated by some English geologists who determined that the dividing line between erosion and deposition along a shore is marked by the number of waves per minute reaching the shore. Essentially the same statement is made by Nichols and Marston since during a storm the waves are longer and fewer reach the shore in a given time.

ORIGIN OF THE LOWS AND BALLS

There are only two possible immediate sources of the sediments that enter into the building of lows and balls. Either the material is picked up and carried out from shore by the waves and currents, or it is picked up from the bottom and redeposited in the balls by the action of the waves. This has led to the consideration of two hypotheses of origin.

First hypothesis.—The material that forms the balls is picked up from the beach and built into ridges by the littoral currents. This explanation was proposed by Gilbert and supported by I. C. Russell. According to Gilbert the material of the ball is composed of "shore drift." This depends largely on the agitation of the water for its continued transportation so it would follow the line of breakers where the greatest agitation takes place rather than following the water margin. This obviously implies a considerable source of material at the point where the current leaves the shore. So far as source and transportation of material is concerned this is the explanation given in the geology textbooks for the building of spits and bars. An examination of spits and bars shows that in most cases they

[32] A. D. Howard, "Hurricane Modification of the Offshore Bar of Long Island," *Geog. Rev.*, Vol. XXIX (1939), p. 407.

[33] A. R. Hunt, "On the Action of Waves on Sea-Beaches and Sea-Bottoms," *Royal Dublin Soc. Proc.* (new ser., 1884), p. 248.

diverge from the shoreline rather abruptly. The explanation itself demands that the littoral current be separated from the shore. This separation may be caused by an abrupt change of direction in the shoreline, or perhaps some obstruction may produce deflection of the shore current, thus causing the spit or bar to be built out in the direction of the movement of the water.

The inner ball sometimes unites with the shore, but when it does the angle formed is very acute and the ball usually remains close to shore for a considerable distance (Fig. 3, *H*). Nowhere has there been observed a tendency, if the shoreline is fairly straight, for the ball to build above the surface of the water as it would if the material were being supplied from the shore as in the building of a spit, although there is usually a projection, or cusp, at the point of junction of the ball with the shore.[34] Also there is very little loss of material from the shore at the point of junction although the ball may extend for a long distance parallel with the shore and so contain a large amount of sediment. The balls out beyond the first one often run parallel with the shore for many miles without having any apparent connection with it. This is clearly shown in the aerial photographs of the Lake Michigan shore. For example, ball No. 2 was traced from the Pentwater piers south for 30 miles to the White Lake piers. Throughout this distance it had no connection in any way with the shore or with the ball in front or back of it.

The littoral currents of this shore are very strong at times and transport considerable sediment. During heavy storms the material stirred up from the bottom by the waves is in sufficient quantity to give a brownish or yellowish color to the water from the shore to and beyond the outer line of breakers. The currents flow south when the wind is from a northerly quarter and north when the wind has a southerly component. Thus, if the balls were formed by littoral currents there would be discontinuation where the uncompleted ridges are being built in from each end like baymouth bars. However, a careful examination by means of aerial photographs of the east shore of Lake Michigan from the southern end to Traverse Bay, a distance of 250 miles, fails to show a single example of this kind.

An examination of the profiles shows that, if a straight edge is laid

[34] Evans, "The Classification and Origin of Beach Cusps," *op. cit.*, p. 618.

as nearly as possible along the bottoms of the lows, the extended line in every case intersects the surface landward of the shoreline. According to Johnson's[35] analysis of offshore-bar profiles this indicates that the material of the balls is thrown up from the lake bottom by the waves and not brought in by the littoral currents. Thus there are five reasons why it is probable that lows and balls are not formed by the work of the littoral currents.

1. Because of the acute angle at which the inner ball usually joins the shore
2. Because the ball does not receive any considerable amount of material from the locality where it joins the shore
3. Because of the failure of the ball to build above the water as does a spit or bar
4. Because of the lack of any considerable connection between the outer balls and the shore, although they are generally higher above the bottom and broader than ball No. 1
5. Because the line of the bottom of the lows prolonged intersects the surface landward rather than seaward of the shoreline

Second hypothesis.—If the material in the balls does not come directly from the shore, then the source of the material is the lake bottom, and the ridge is built up by the work of the incoming waves. This is essentially the same as the explanation offered by De Beaumont[36] in accounting for the origin of offshore bars. Johnson[37] supports De Beaumont's hypothesis and goes into an extensive analysis of the method so far as it applies to the building of offshore bars. However, later in his book he discusses lows and balls[38] and quotes Gilbert and I. C. Russell freely but does not quote De Beaumont. Regarding this seeming anomaly Johnson says: "When I wrote my *Shore Processes* I was not absolutely sure the 'low and ball' is the forerunner of the offshore bar although I believed that to be the most probable explanation."[39]

If lows and balls are built by the work of the onshore waves, it is necessary to call attention here to certain features of wave action of especial importance in explaining the process. In waves of oscillation the water particles have a circular motion. As wave after wave passes, each particle continues to describe a circular movement and comes back to its original position, its total movement of translation

[35] *Op. cit.*, pp. 356–67.

[36] Elie de Beaumont, *Leçons de géologie practique* (Paris, 1845), pp. 223–352.

[37] *Op. cit.*, pp. 350–67. [38] *Ibid.*, pp. 486–89. [39] Private communication.

being zero. However, wind causes surface layers of water to move slowly to leeward and thus the water particles, returning in their orbits to the upper part of the wave, are slightly to leeward of their former positions. This translation movement becomes rapidly less with depth. In studying the origin of lows and balls the circular motion of the water particles is very important.

A wave of translation is the kind of wave formed when a solid body is suddenly plunged into water. The mass of water in the wave moves forward. Such a wave is not accompanied by a trough and there is no return movement of the water such as occurs in a wave of oscillation. Such waves are very efficient in the transportation of material to leeward and are the chief factor in preventing the ball from building above the water surface.

It has been stated that waves of oscillation may be changed into, or give rise to, waves of translation and that the reverse may also happen. Johnson says: "This type of wave (translation) is extensively developed in the shallow waters along coasts, the waves of oscillation generated in deep water frequently become more or less completely transformed into waves of translation as they reach the shore."[40] This is in accordance with the arguments of J. Scott Russell,[41] who is generally credited with being the discoverer of the wave of translation. However, Hunt[42] states that waves of oscillation are not changed into waves of translation although they do give rise to currents of translation. So far as transportation of sediments is concerned it makes little difference whether waves of oscillation give rise to waves of translation or set up currents of translation.

Careful observations of the movements of water as waves of oscillation break over a shallow bar or in the shallows near shore show that following the breaking of the wave the water in the upper part of the breaker is thrown forward as a ridge of water with a foaming, gliding front. This ridge travels rapidly to leeward across the bar and carries any floating object with it. The water between these ridges may be either stationary, moving forward, or an oscillatory movement may be present on the bottom, depending on the depth of the water. The result is a more or less rapid forward movement of material.

[40] *Op. cit.*, p. 33. [41] *Op. cit.*, pp. 311–90. [42] *Op. cit.*, p. 359.

THE ORIGIN AND BEHAVIOR OF BREAKERS

Johnson describes the breaking of oscillatory waves as follows:

> The size of the orbital paths increases as waves enter shallower water, while at the same time the volume of water is decreasing. With constantly enlarging orbits and diminishing water supply, there must come a time when the volume of water is insufficient to build up the entire wave form, the deficiency manifesting itself as a "hollowing" of the front of the wave. The water available endeavors to curve through the entire orbit, but on reaching the top of the circle finds itself unsupported and collapses.[43]

MacClintock, in discussing the breaking of waves, says: "When the increasing orbital velocity of the upper wave particles becomes greater than the forward motion of the wave, the wave breaks."[44] Careful observation by the writer shows there are two stages in the formation of breakers, or perhaps it is allowable to say there are two kinds of breakers. These may be designated as the "gliding" type and the "plunging" type (Figs. 4, 5, and 6).

In the gliding breaker there is merely a steepening of the front of the wave sufficient to cause the water in the front of it to move rapidly downward and forward under the influence of gravity. Considerable turbulence of the surface results and some air becomes mixed with the water of the wave front. This gives it a characteristic light-gray or white appearance. There are several ways in which this type of breaker is originated. One form of it is the ordinary whitecap. This is frequently produced by a rising wind which pushes forward and oversteepens the front of the wave. A very similar effect occurs when a wave advances into shallowing water but with insufficient velocity to build up into the plunging type. This can often be observed when smaller waves move over a shallows where the larger waves are being forced to plunge. The gliding breaker may also result from the coincidence of two wave crests when two or more wave systems are running simultaneously. When the longitudinal axis of the superimposed smaller wave is to the leeward of the axis of the larger one a steepening of the wave front results that may cause a downward gliding of the water in it.

Waves produced inshore from plunging breakers in shallow water

[43] *Op. cit.*, p. 16.

[44] Paul MacClintock, "The Breaking of Waves in Shallow Water," *Jour. Geol.*, Vol. XXXII (1924), pp. 407–9.

are usually of the gliding type. It is not certain they are true waves of translation, but they do possess many of the characteristics of

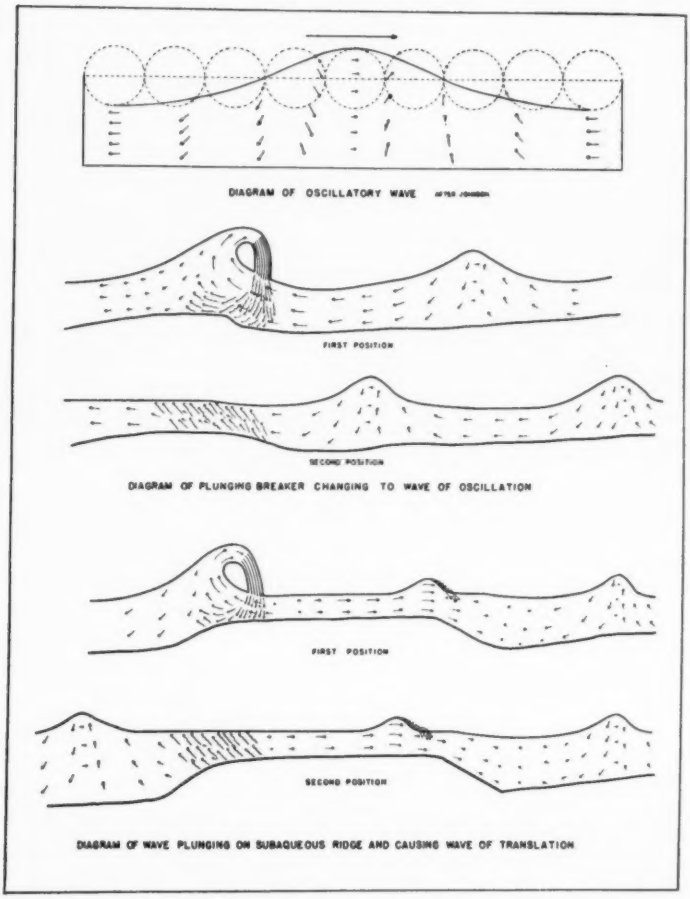
DIAGRAM OF OSCILLATORY WAVE
FIRST POSITION
SECOND POSITION
DIAGRAM OF PLUNGING BREAKER CHANGING TO WAVE OF OSCILLATION
FIRST POSITION
SECOND POSITION
DIAGRAM OF WAVE PLUNGING ON SUBAQUEOUS RIDGE AND CAUSING WAVE OF TRANSLATION

FIG. 4.—Diagram of breakers

such waves and are frequently important in causing a rapid forward movement of material.

The bottom below the gliding breaker is not much disturbed by its passage. In deep water the effect is little more than the recipro-

FIG. 5.—Plunging breaker

FIG. 6.—Turmoil after breaking

cating surge produced by an unbroken wave of oscillation. In shallow water this backward and forward surge is increased and the resulting motion sometimes throws the bottom sediments into suspension. They are then subject to movement by any currents that may be present. Obviously where the gliding breaker has some of the characteristics of a wave of translation the movement of these sediments will, in general, be forward in the direction of movement of the wave.

In the plunging type of breaker the water in the wave crest curls over and falls freely and nearly vertically as does the water in a waterfall. Such breakers generally occur where the depth of the water is approximately equal to the height of the waves. The downward plunging water mass delivers a blow on the bottom and some of this energy is expended in moving any loose sediment that may be present. The violently disturbed material is all, or nearly all, carried to windward, that is, toward the axis of the breaking wave by the orbital motion just in front and below the breaking crest. This is the outward motion of the trough of the oscillation wave which is drawn close in under the crest of the breaker because of the distortion of the wave at the time of breaking (Fig. 4). This movement is often strong enough to erode the bottom and cause sand to be drawn up into the breaker orbit before the bottom is disturbed by the plunge.[45] This movement and its effects were also observed by Wheeler.[46]

[45] On August 24, 1939, I stood on the north pier at the entrance of White Lake and watched the action of the breakers on the first ball as waves about 3 feet high came in toward the shore with their axes almost perpendicular to the pier. Just before and during the breaking of each wave the sand, which was clearly visible in the clear water, was seen to be picked up and moved backward and forward. Often some of the sand could be seen to be carried in the back part of the breaker nearly to the top, and occasionally it came clear over and was carried down again in the water of the plunging crest. After the passing of the breaker a turmoil of the water continued, and in various places it "boiled" and came rapidly to the surface, bringing sand with it (Fig. 6). These violent movements are caused, at least in part, by escaping air which has been entrapped and carried down during the plunge of the breaking crest. Thus, after the passing of the breaker, sand could be seen in suspension for some time. The turmoil and suspension sometimes continued until a part of the sand was again caught up in the next breaker. When an oscillation wave passed over without breaking, the suspended sand appeared to merely rise and fall. Careful observations of the suspended sediment showed that, after being thrown into suspension, it was seldom or never carried toward shore but either moved outward or settled approximately in the same place where it was first observed.

[46] *Op. cit.*, p. 117.

Offshore movement of sediment raised by the plunging breakers is the fundamental cause for the beginning and maintenance of the low and ball. As waves come into shallowing water and reach the depth at which they are forced to break, sand on the bottom is moved outward from shore (Fig. 4). As the breakers succeed each other this motion is continued until a trough is cut in the sloping bottom and the material from it is piled some little distance outward.

A factor which is probably of some importance in maintaining the low and ball after the ridge is once formed is the induction or reverse eddy current which is set up in the troughs as a result of the shoreward moving water at the surface. Eddy currents of this kind flowing in a direction opposite to the primary currents and dependent on them for their existence have been called induction currents by Cornish.[47] Horizontal induction eddies are very commonly observed along the shores of rivers and lakes but vertical induction eddies are not so well known since they occur beneath the surface. Such currents were observed by Evans[48] between parallel ridges on a broad terrace of Crystal Lake near Frankfort, Michigan, during an onshore wind. A glance at a section through a series of lows and balls of Lake Michigan (Fig. 2) shows that the conditions are ideal for the occurrence of such currents. This current might be considered a part of what is commonly referred to as the undertow although it is not a hydraulic current, as the undertow is generally assumed to be, but owes its existence to friction with the overlying water which is moving toward the shore. Such a current is evidently an appreciable factor during storms in carrying the suspended sediments from the trough bottoms to the windward-lying balls.

Each low and ball marks the average breaking position of waves of a certain size. Table 2 shows an increase in depth of 4 feet over successive balls in passing from the shore outward and an increase of 6 feet in the depths of the lows. It is evident that the outer low and ball is formed by larger storm waves while the inner ones are the work of the smaller waves.

During heavy storms the waves produce plunging breakers on all balls. The larger waves plunge on the outer ball and after breaking cause the formation of smaller oscillation waves to leeward, some of

[47] *Op. cit.*, pp. 529–30.

[48] "The Undertow," *Science*, Vol. LXXXVIII (new ser., 1938), p. 281.

which break on the next inner ball. This continues until those waves which finally reach the shore are much reduced in size. In times of moderate winds the waves are of less size and pass over the outer balls and do not plunge until they reach comparatively shallow water over the inner balls. Thus it is probable that since the stronger winds are of relatively short duration, the inner balls are disturbed and worked over more than the outer ones.

CONCLUSIONS

1. There are two types of breakers: the plunging type and the gliding type. The plunging type stirs up sediment from the bottom and moves most of it to windward where it may deposit and form a ridge; the gliding type has a less steep front and the water in this front moves down and forward in a direction opposite to that of the movement of the particles in the wave orbit but does not disturb the bottom to any extent.

2. The low with the ball to the windward of it constitutes a unit. It is built by a line of plunging breakers whose position is determined by the depth of the water and the size and velocity of the oscillation waves that produce it.

3. Where the slope of the bottom is so related to the size of the waves that several lines of plunging breakers are produced there will also be a series of lows and balls, each line of plunging breakers producing a low and ball.

4. Lows and balls do not travel shoreward with the waves but remain stationary except for readjustments due to changing conditions in the size and velocity of the waves.

5. Lows and balls are subject to adjustments with respect to depth of water, size of waves, velocity of waves, and currents. This adjustment may result in change of shape or size of the low and ball or its migration either away from shore or toward it.

6. Lows and balls are generally larger with distance from the shore. This is because the offshore breakers are larger and carry more energy which at the time of plunging is partly expended in moving bottom sediments.

7. The bottoms of the lows and the crests of the balls occur at increasing depths with distance offshore and the bottoms of the lows deepen faster than the crests.

8. Oscillation waves which produce plunging breakers of a size to affect the inner lows and balls may pass over the outer balls without breaking. Such waves have little or no effect on the outer lows and balls.

9. If a fall in water-level occurs, the inner ball will begin migrating toward shore as soon as its top comes within the sphere of action of the currents or waves of translation formed by the plunging wave crest. As soon as this occurs the ball takes on the characteristics of a subaqueous dune.

10. When a considerable rise in water-level occurs, the lows and balls already formed do not migrate toward shore. Instead a new low and ball is formed inshore from the old series. Should the rise be adequate and permanent, the farthest offshore low and ball then has so great a depth of water over it as to be out of the sphere of influence of the largest waves and thus becomes a fossil low and ball.

11. On the inshore terraces of the eastern shore of Lake Michigan ridges form in various ways which sometimes have a close resemblance to lows and balls. These move freely under the influence of waves and currents. The material in these ridges may be distributed again over the terrace but frequently they migrate shoreward during times of moderate winds and unite with the beach ridges.

12. Along the eastern shore of Lake Michigan structures such as piers, which project several hundred feet out into the lake nearly perpendicular to the shore, interfere with and cause an irregular arrangement of the lows and balls for a distance of at least half a mile each way along the coast.

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THRUST FAULTING OF GRENVILLE GNEISSES NORTHWESTWARD AGAINST THE MISTASSINI SERIES OF MISTASSINI LAKE, QUEBEC

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Geological Survey, Canada

ABSTRACT

In the Mistassini Lake area the Timiskaming and the Grenville subprovinces of the Canadian Shield meet along a faulted contact extending northeast. The Grenville gneisses along the faulted contact are crushed and locally mylonitized and are believed to have been thrust northwestward against the rocks of the Timiskaming subprovince. The late pre-Cambrian Mistassini series, northwest of the gneisses and consisting of dolomite overlain by an iron formation, dips gently southeast toward the faulted contact along which the rocks are much disturbed.

INTRODUCTION

Mistassini Lake (Fig. 1) just beyond the Hudson Bay–St. Lawrence divide, 300 miles northwest of Quebec city, is one of the several places in the Canadian Shield east of Hudson Bay in which late pre-Cambrian clastic, carbonate, and iron-rich sediments occur. Apart from investigations to assess the value of the iron formations of this region, our knowledge of the stratigraphy, structure, and distribution of these rocks rests mainly on surveys by A. P. Low.[1] Mistassini Lake area was one of the first of these areas of late pre-Cambrian to be investigated, but at the time of its exploration little headway had been made in dividing the Canadian Shield into subprovinces characterized by distinctive rock assemblages. Special interest now centers on the area owing to its location at the boundary of two of the subprovinces of the shield—the Timiskaming and the Grenville—and the light which it throws on their relationships. Furthermore, the pre-Cambrian age of the Mistassini strata has been challenged by some, but information now available should dispel any doubt as to the age and general correlation of these rocks. The present purpose is to describe the two rock assemblages in the Mistassini area and their faulted contact, as determined by a six weeks' study of the region in 1938.

[1] Extracts from reports on the District of Ungava, *Dept. of Mines and Fisheries, Bureau of Mines Quebec* (1929).

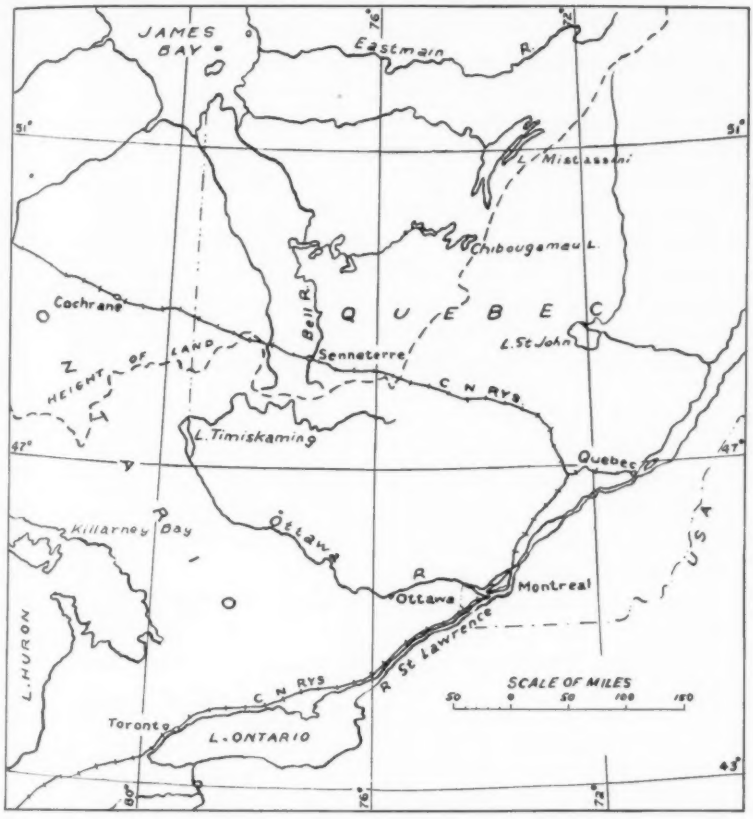

FIG. 1.—Index map showing location of Mistassini Lake

PREVIOUS WORK

Mistassini geology has been briefly discussed by James Richardson,[2] Walter McOuat,[3] A. P. Low,[4] and A. E. Barlow.[5] Richardson merely reported that flat-lying limestones, containing what he believed to be an orthoceratite and a coral, occurred at Lake Mistassini. McOuat partly surveyed Mistassini Lake, and Low completed the survey and described the strata which he correlated with the Cambrian, in accordance with the usage of his time, to imply their resemblance to the Animikie rocks of the Lake Superior region. Low considered the strata only slightly deformed and concluded that the Mistassini Lake basin existed before the sediments were deposited. Barlow, the only other geologist to visit the lake, perhaps influenced by the gentle dip and the presence of what he believed to be organic remains, regarded the Mistassini strata as probably Ordovician.

PRE-CAMBRIAN SUCCESSION OF THE MISTASSINI REGION

The boundary between the Grenville and the northeast end of the Timiskaming subprovince extends northeastward across the Mistassini region 7–14 miles southeast of Mistassini Lake. The great difference between the rocks of these two subprovinces in the Mistassini area is shown in Table 1.

The two groups which play the most important part in the structure of the Mistassini district are the Mistassini series of Mistassini Lake and the gneisses of the Grenville subprovince southeast of this lake.

TIMISKAMING SUBPROVINCE

In the Mistassini area the early pre-Cambrian rocks of this subprovince consist of red and gray granitic types, in part gneissic, with minor amounts of hornblende schist. Southwestward toward

[2] "Report of the Country North of Lake St. John," *Geol. Surv., Canada, Rept. of Progress 1870–71*, pp. 294–95.

[3] "Report of Explorations of Country between Lake St. John and Lake Mistassini," *Geol. Surv., Canada, Rept. of Progress 1871–72*, pp. 115–19.

[4] "Report of the Mistassini Expedition," *Geol. Surv., Canada, Ann. Rept.*, Vol. I (1884–85), pp. 1–55*d*; *Geol. Surv., Canada, Ann. Rept.*, Vol. VIII (1895), pp. 196*L* and 266*L*.

[5] "Geology and Mineral Resources of the Chibougamau Region, Quebec," *Dept. of Colonization, Mines, and Fisheries, Quebec, Mines Branch* (1911), pp. 131–33.

Chibougamau Lake[6] early pre-Cambrian granitic rocks decrease in amount, and a great variety of lavas and sediments, of Keewatin and Timiskaming types, and basic intrusives occur.

Two late pre-Cambrian series, which rest unconformably on early pre-Cambrian rocks, are found in the Mistassini-Chibougamau district. The Chibougamau series,[7] the older of the two, is not seen to underlie the Mistassini series. Nevertheless, the regional dis-

TABLE 1

	Timiskaming Subprovince	Grenville Subprovince
Late Pre-Cambrian..	Mistassini Series Iron formation, ferruginous chert, dolomite, and dolomitic limestone	
	Unconformity	
	Chibougamau Series Conglomerate, arkose, banded graywacke	Granite gneiss, biotite, and hornblende gneiss
	Unconformity	
Early Pre-Cambrian..	Granite, syenite, hornblende schist	

tribution of the two series shows fairly clearly that the Chibougamau is the older and that it was partly eroded before the Mistassini was deposited. The Chibougamau series occurs in small faulted blocks at the northeast end and north of Chibougamau Lake (Fig. 1) and at Wakonichi Lake (Fig. 2). A few small outcrops of arkose and conglomerate probably belonging to this series are seen near the southwest end of Mistassini Lake, in the faulted zone between Mistassini strata and the gneisses to the southeast, and immediately northeast of Mistassini strata on Temiscamie River. The series con-

[6] G. W. H. Norman *et al.*, "Chibougamau Sheets," *Maps 397A and 398A, Geol. Surv., Canada* (1938).

[7] J. B. Mawdsley and G. W. H. Norman, "Chibougamau Lake Map-Area, Quebec," *Geol. Surv., Canada, Mem. 185* (1935), p. 43.

sists of conglomerate and arkose with local varved deposits of silt grade containing scattered pebbles, cobbles, feldspar fragments, and some shale. It dips gently, except near faults, and rests with marked

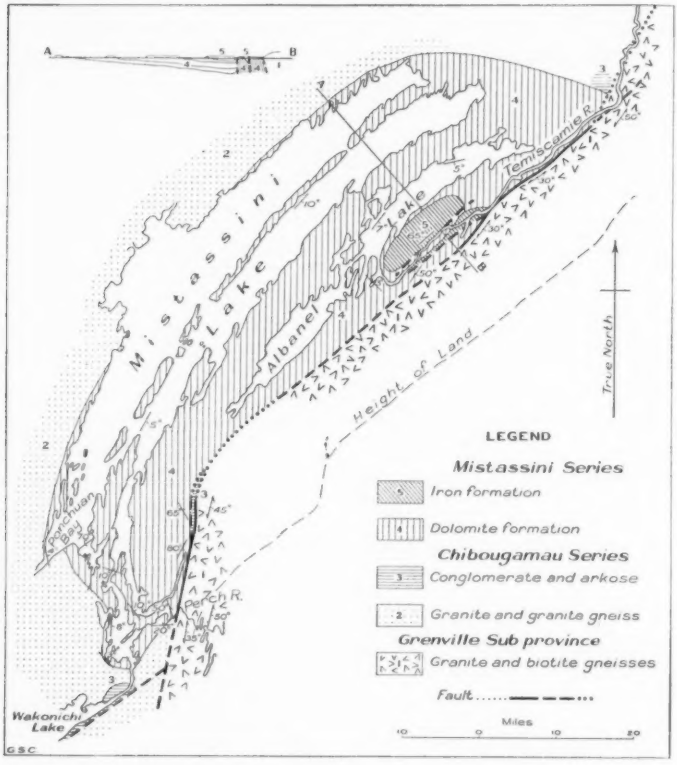

FIG. 2.—Sketch map showing geology of Mistassini Lake region

unconformity on early pre-Cambrian rocks. The unconformity is well exposed at the north end of Wakonichi Lake where granite passes gradually into residual arkose and breccia, and this material which is about 40 feet thick in turn passes gradually into conglomerate composed almost entirely of granite debris. The peculiar lithology of the series, particularly of the varved beds of possible glacial

origin, and the great unconformity that separates the series from older rocks point to its possible correlation with part of the Cobalt series of the Timiskaming Lake region,[8] Quebec.

The Mistassini series may be divided lithologically into two formations, the lower of well-bedded to massive dolomites and dolomitic limestones[9] and an upper one of ferruginous cherts, dark slaty shale, and jaspilite types composed of thin alternate layers of magnetite and silica. The dolomite formation underlies Mistassini and Albanel lakes and a strip along Temiscamie River. The iron formation is restricted to a comparatively small area east of the center of Lake Albanel.

DOLOMITE FORMATION

There are no continuous sections of this formation showing its thickness or the complete succession of its component beds. It is possible, however, to divide the formation into three members, of which the basal member (estimated to be between 200 and 300 feet thick) consists of black, slaty, finely bedded dolomites, the middle member (between 200 and 300 feet thick) of well-bedded to massive white, gray, to buff-weathering dolomite, and the uppermost member (100–150 feet thick) of cryptozoon- and sand-bearing dolomites.

The basal member fringes the southwest end of the Mistassini basin and probably underlies the wide part of the lake northwest of the central chain of islands. The strata of this member (Fig. 3) break easily into platy layers, 1 inch or less thick, or in more massive beds show fine bedding laminations that in places are cross-bedded. They weather buff but on fracture are dark gray to black. Ferruginous, siliceous, and black argillaceous impurities are present in these beds, and variations in composition are clearly brought out in places by differential weathering into small alternate ridges and depressions. At Ponchuan Bay the basal dolomitic beds of this member rest directly on gray granite gneiss. A few angular blocks of the gneiss and small discontinuous seams of sandy composition occur

[8] M. E. Wilson, "Timiskaming County, Quebec," *Geol. Surv., Canada, Mem. 103* (1918), pp. 103–13.

[9] Carbonaceous shale may be interbedded with the dolomites and occurs as erratics in the Mistassini region.

in the dolomite above the contact, and irregular seams of dolomite, produced probably by the infiltration of limy muds, are found in fractures in the underlying gneiss. A cryptozoon horizon 2–3 feet thick occurs 25–50 feet above the base of the member at the southwestern part of the basin. The cryptozoon structures are closely packed together and in horizontal section are 1–2 inches in diameter. In vertical cross section they are rudely triangular with apex downward, somewhat like oysters 2–3 inches long in appearance. They are divided by numerous septa that are convex upward with broad flat crests and small steep sides, and some send out offshoots as if by budding.

The more resistant parts of the middle member form the islands along the center of Mistassini Lake and underlie the long strip of land separating this lake from Lake Albanel. In the southwestern portion of the basin they swing around with the structure and underlie the narrow strip of land between the two southern bays of Mistassini Lake to come in faulted contact with arkose, conglomerate, and mylonitized granite north of Perch River. The exposed parts of this member (Fig. 4) consist of finely to quite visibly crystalline dolomite, varying in color from almost white to gray and bluish gray. They occur in beds a few inches to 10 feet or more thick, ranging from fairly pure dolomite to impure varieties with siliceous or ferruginous impurities.

The uppermost member of the formation is exposed on the shores and islands of Lake Albanel. In bedding, color, and composition the beds of this member resemble those of the middle member. They contain, however, numerous concentric cryptozoon structures whose inner detail has been obliterated by recrystallization. The structures are either circular, elliptical, or irregular in cross section and range from 6 inches to 10 feet or more in diameter, their size depending on the closeness of spacing. In vertical cross section they seem to increase in size upward. The concentric bands consist of comparatively pure pale-gray carbonate, $\frac{1}{8}$–$\frac{1}{2}$ inch wide, alternating with dark bands, $\frac{1}{16}$ inch in width, and contain scattered sand grains. Many of the beds in the uppermost member, particularly those with cryptozoon structures, contain up to 25 per cent of remarkably well-rounded sand grains, 1 mm. or less in diameter, and in places are much cross-bedded. The sand grains have frosted surfaces like

FIG. 3.—Overturned well-bedded slaty dolomite on Temiscamie River

FIG. 4.—Gently folded massive gray dolomite at the narrow southwest part of Mistassini Lake.

wind-blown material. Pisolitic structures are present also in some of the beds and occur in small irregular "reefs."

IRON FORMATION

The iron formation outcrops on a few islands in Lake Albanel at the mouth of Temiscamie River, along a northwesterly facing cuesta between Lake Albanel and Temiscamie River and at a few places along Temiscamie River. Its thickness is probably between 100 and 200 feet. The complete succession of the formation is not exposed nor is its contact with the underlying dolomites. The lower part of the formation consists of rusty-weathering ferruginous chert consisting of very fine-grained quartz and ferruginous carbonate, probably siderite, intimately mixed together in nearly equal proportions. The quartz grains range from 0.0007 to 0.15 mm. in size and the carbonate grains from 0.15 to 0.5 mm. A few small wisps of ferruginous chlorite (chronstedite?) of high birefringence and yellow-green to nearly black pleochroism occur in the rock together with some limonite. The strata exposed between Lake Albanel and Temiscamie River are ferruginous cherts containing 10–15 per cent magnetite in places; there the carbonate is almost entirely altered to limonite. Along Temiscamie River cherty iron-carbonate rocks crop out at a few places, but the characteristic rocks of the formation there are dark slate, ferruginous slate, and beds consisting of interbanded magnetite and cherty silica in layers $\frac{1}{2}$ inch or less to 1 inch or slightly more in thickness. Some of the beds contain nearly 50 per cent magnetite chiefly in rounded or slightly irregular granules in a cherty matrix. The granules are 1 mm. or less in size and may originally have been greenalite. Beds of iron ore, if present, are not exposed, nor is their presence indicated by glacial erratics or loose fragments in the drift. There are boulders in the drift at one place along the river that consist of interbanded magnetite and chert cut by crocidolite (fibrous blue asbestos) veins $\frac{1}{4}$ inch or so in width. The crocidolite in these veins closely resembles specimens from the South African deposits.

AGE OF THE MISTASSINI SERIES

In lithology and stratigraphic succession the Mistassini series, consisting of an iron formation overlying dolomite beds, closely re-

sembles the other iron formations and associated strata in the region east of and about Hudson Bay.[10] These iron formations and associated strata are more like the Animikie of the Lake Superior region than any other series and are customarily considered late pre-Cambrian in age.

GRENVILLE SUBPROVINCE

In the Mistassini region the characteristic rocks of this subprovince are medium-grained gray biotite gneiss and pink gneissic granite. In places hornblende-zoisite gneiss, derived possibly from gabbro, biotite schist, and altered diabase or gabbro occur. In contrast to the early pre-Cambrian rocks underlying the Mistassini strata these rocks are regionally metamorphosed. They cannot be traced into less-altered types, and their age, relative to the Mistassini strata, is difficult to determine owing to faulting. Probably the pink granite gneisses which are younger than the gray gneisses are older than the Mistassini strata. The question, however, of the age of these rocks relative to the various pre-Mistassini rocks of the Timiskaming subprovince is still an open one.

STRUCTURE OF THE MISTASSINI REGION

The strata of Mistassini and Albanel lakes, as shown in Figure 2, form an arcuate basin-like structure that is cut off sharply to the southeast by upthrust gneisses of the Grenville subprovince. Everywhere except along the southeast edge of the basin the strata dip gently, for the most part southeastward, and rest unconformably on early pre-Cambrian granitic rocks. The strata along the southeast edge of the basin, however, are much crumpled (Fig. 5), folded, and faulted near the fault boundary with the Grenville gneisses. The gneisses extend southeastward from Mistassini area to the Paleozoic rocks of the St. Lawrence lowlands.

The arcuate structure of Mistassini and Albanel lakes, except along its southeast edge, dips in general from 5°–10° southeast,

[10] C. K. Leith, "An Algonkian Basin in Hudson Bay," *Econ. Geol.*, Vol. V (1910), pp. 227–46; E. S. Moore, "The Iron Formation on Belcher Islands, Hudson Bay, with Special Reference to Its Origin and Its Associated Algal Limestones," *Jour. Geol.*, Vol. XXVI (1918), pp. 412–38; J. E. Hawley, "Geology and Economic Possibilities of Sutton Lake Area, District of Patricia," *Ont. Dept. Mines*, Vol. XXXIV, Part 7 (1925), pp. 1–56.

though locally the strata are horizontal and in places dip toward the northwest. Low states that thrust faults occur along the central chain of islands and along the southeast side of Mistassini Lake. His evidence, the subdued cuesta topography at these places, is not very convincing. It seems unlikely that the gently dipping strata of Mistassini and Albanel lakes are much faulted, although on one small island in Ponchuan Bay strata striking 55° east of north are overturned northwestward. The strike of the overturned strata is parallel to the straight south shore of near-by Ponchuan Bay, along which the dolomite-granite contact is apparently shifted to the left as if by faulting.

Along the southeast edge of the arcuate structure against the upfaulted gneisses to the southeast the Mistassini strata are much disturbed in a zone 1–4 miles wide. The change from the gentle dips, which characterize the main part of the arcuate structure, to the steep dips of strata in the disturbed zone takes place rather abruptly at the few places where the change was observed. The disturbed zone is widest southeast of Albanel Lake, where the greatest thickness of Mistassini strata is present, and narrow at either end of the Mistassini basin, where the total thickness of strata is small. At the widest part of the disturbed zone the strata are folded and faulted as diagrammatically shown in Figure 2. In the narrow parts of the disturbed zone the strata appear to be only titled on edge or overturned. There is, therefore, a relationship between the width of the disturbed zone, the thickness of strata present, and the amount of folding and faulting that the strata have undergone. Perhaps the explanation of this relationship is that the disturbed zone dips southeast under northwesterly overthrust gneisses, and that at either end of the basin the thin sections of strata now exposed were too well protected by underlying crystalline rocks to be folded.

The folds and faults in the wide part of the disturbed zone parallel the main northeasterly fault along which the gneisses are thrust against the Mistassini strata. They are interpreted as subsidiary structures produced in the Mistassini strata by the thrust faulting of the gneisses toward the Mistassini basin. The subsidiary faults are more numerous than is shown in Figure 2, but the writer lacked sufficient time to map them in detail or to find out whether they have the imbricated pattern present in the Northwest Highlands of

Scotland, where gneisses and schists are thrust over late pre-Cambrian and early Paleozoic rocks.[11] The faults are marked in some cases at least by breccia zones in which the dolomite is fragmented (Fig. 6). One of these fault-breccia zones, along Temiscamie River, is almost completely replaced by cherty quartz and stands out as a snow-white ridge 250 feet wide and $\frac{1}{2}$ mile long, like the giant quartz veins[12] of the Northwest Territories of Canada.

The faulted contact of the Grenville gneisses with Mistassini strata extends, as shown in Figure 2, 10° east of north in the southwest part and 55° east of north in the northeast part of the basin. In the intervening section near the southwest end of Albanel Lake the contact was not examined but must make a fairly sharp bend of 45°. The sharp bend suggests perhaps that two faults intersect or diverge from one another at this point and that the contact occurs along a series of faults rather than along one great rupture, particularly as throughout the region 150 miles southwest of Mistassini Lake two series of faults, one approximately 10°, the other 45°, east of north, occur. It should be noted in this connection that the straight south shore of Ponchuan Bay, which is most probably a fault line, is in direct alignment with the faulted boundary of the gneisses southeast of Albanel Lake. It may be, therefore, that two faults meet near the southwest end of Albanel Lake and that the fault striking 55° east of north extends southwest across the Mistassini basin.

Little evidence was collected regarding the dip of the fault or faults along which the gneisses are brought up against the Mistassini strata. At one high hill capped by gneisses and schists near the north end of Lake Albanel, dolomite strata dip southeast at the base of the hill as if to underlie the gneisses and schists above. The contact is concealed there, however, in a 900-foot interval of no exposures. Both the dip of the Mistassini strata and of the foliation and banding of the gneisses and schists near the faults are to the southeast, suggesting that the faults also probably dip in this direction. The dip of the Mistassini strata nearest the faults ranges from 50° to 80°, and that of the gneisses and schists from 25° to 65°.

[11] B. N. Peach, John Horne, *et al.*, "The Geological Structure of the Northwest Highlands of Scotland," *Geol. Surv., Great Britain* (1907).

[12] D. F. Kidd, "Great Bear Coppermine River Area, McKenzie District, Northwest Territories," *Geol. Surv., Canada Summ. Rept.*, Part C (1931), p. 54c.

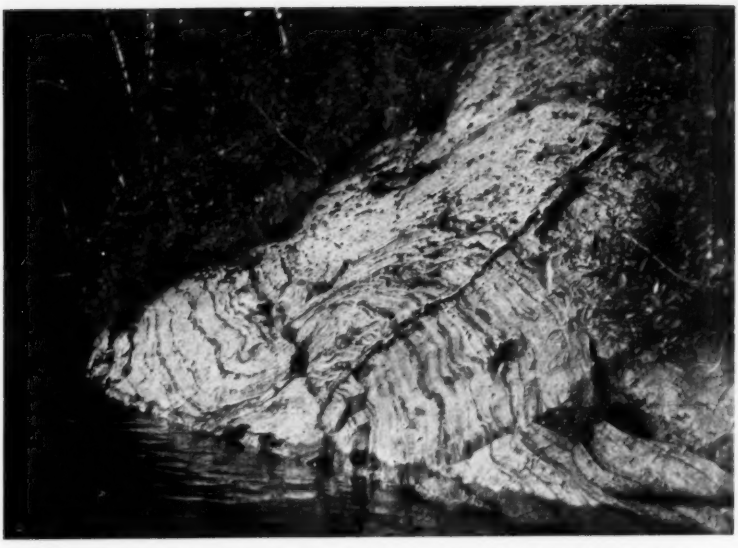

FIG. 5.—Drag-folded, steeply dipping dolomite in the disturbed belt along Temiscamie River.

FIG. 6.—Brecciated dolomite, Temiscamie River

Immediately northwest of the faults the Mistassini strata are intensely deformed and in places faulted and brecciated; likewise the gneisses along the southeast side of the faults are intensely crushed and in places severely mylonitized. The zone of crushed gneisses is $\frac{1}{2}$ mile wide, and it is difficult to determine the original character of many of its rocks. Rocks believed to have been originally granite are crushed to a fine mosaic of epidote, white mica, albite, microcline, and quartz; other mylonitized rocks are obviously derived from, and can be traced into, coarse granite gneiss. The gray gneiss in the zone of crushing is sheared out to resemble schist or is so smashed that the original banding is hard to detect. Away from the crush zone the gray gneiss contains calcic oligoclase, quartz, and biotite; in the crush zone the plagioclase is albite accompanied by much epidote, and biotite is partly or completely altered to chlorite and amphibole.

THE POSSIBLE EXTENT OF THE MISTASSINI FAULT ZONE

In 1936 the writer[13] advanced the idea that a zone of faulting extended northeastward across Ontario and Quebec from Killarney Bay, Lake Huron (Fig. 1). The evidence for this was the fairly abrupt southeastward termination of the Huronian formations north of Lake Huron and of the Keewatin and Timiskaming rocks in Quebec against a region of gneisses called by many the Grenville subprovince. In the Chibougamau district of Quebec in particular, the evidence supporting the idea of a regional fault zone—numerous large northeast faults and diabase dikes that occur in a zone 24 miles wide along the northwest side of the Grenville gneisses—is strong. The fault zone separating the Grenville gneisses from the Mistassini strata is a northeast continuation of this postulated zone of faulting across Ontario and Quebec. The Mistassini part of this zone has a length of more than 100 miles, and the latest movements there are clearly younger than the Mistassini strata. Further work, however, will be needed to bring out the extent to which faulting, as seen in the Mistassini area, occurs generally along the northwest boundary of the Grenville subprovince of the Canadian Shield, as well as to determine the age of this faulting.

[13] G. W. H. Norman, "The Northeast Trend of Late Pre-Cambrian Tectonic Features in the Chibougamau District, Quebec," *Trans. Roy. Soc. Can. Sec. IV*, Vol. XXX (1936), pp. 119–28.

A RECORD OF *LYONOTHAMNUS* IN DEATH VALLEY, CALIFORNIA

DANIEL I. AXELROD[1]

ABSTRACT

The presence of *Lyonothamnus* in the Furnace Creek formation (undescribed) is discussed from the standpoint of plant distribution in arid regions. It is suggested that the Catalina ironwood may have lived in canyons adjacent to the Furnace Creek basin during late Miocene or early Pliocene time, when rainfall in these areas is considered to have ranged from 13 to 15 inches annually.

The discovery of an identifiable fossil plant in the Tertiary sediments of Death Valley adds some important information to the little-known paleobotanical record of the desert region. The material was collected at a locality discovered by Mr. H. Donald Curry during geological work in the region. The writer visited the area with Curry in the spring of 1939 while he was engaged in later Tertiary paleobotanical field work in the Great Basin area under the auspices of the Carnegie Institution of Washington. Mr. James F. Ashley, who served as field assistant, discovered the fragment.

Although the present record comprises only a lateral leaf lobe, it is sufficiently diagnostic to be referred without hesitation to the fossil species *Lyonothamnus mohavensis* Axelrod.[2] Its nearest living relative, the Catalina ironwood (*Lyonothamnus floribundus* Gray), is a medium-sized tree monotypic of the genus, now endemic to the Channel Islands off the coast of southern California. In addition to its Furnace Creek occurrence, *Lyonothamnus* has been discovered at five other localities in California and Nevada during the last few years (Fig. 1).

The specimen comes from the upper portion of the Furnace Creek formation (undescribed), where it occurs together with some indeterminable grasslike or reedlike plants which were noted first by

[1] National Research Fellow, 1939–40, engaged in research at the United States National Museum.

[2] D. I. Axelrod, "A Miocene Flora from the Western Border of the Mohave Desert," *Carnegie Inst. Wash. Pub. No. 516* (1939), pp. 107–8.

Spurr in 1903.[3] The outstanding features of these beds, which are well exposed in Furnace Creek Wash several miles southwest of Furnace Creek Inn, have been summarized by Curry:

FIG. 1.—Sketch map showing the past and present distribution of *Lyonothamnus* in California and Nevada. *X* = Furnace Creek (Mio-Pliocene); *1* = Tehachapi (Middle Miocene); *2* = Mint Canyon (Upper Miocene); *3* = Upper Modelo (Upper Miocene); *4* = Coal Valley (Upper Miocene); *5* = Mulholland (Middle Pliocene); *R* = Recent distribution (Channel Islands).

The Furnace Creek formation consists of several thousand feet of lavas, pyroclastics and terrestrial sediments lying in the area drained by Furnace Creek Wash in the central Death Valley region. The lower part of the formation is made up primarily of highly colored volcanics and pyroclastics, with some fanglomerates, all presumably in unconformable contact with earlier Tertiary

[3] J. E. Spurr, "Descriptive Geology of Nevada South of the 40th Parallel and Adjacent Portions of California," *U.S. Geol. Surv. Bull. 208* (1903), p. 189.

rocks. The pyroclastics grade laterally and vertically into tuffaceous lake and playa sediments. The upper part of the formation is typically a gray or cream tuffaceous playa clay that weathers tan or golden yellow. Tuffaceous sandstones and freshwater limestones of similar color are common also. A younger formation (Ryan, undescribed) unconformably overlies the Furnace Creek in some areas. The whole assemblage apparently represents an accumulation in an arid intermontane basin.

The beds have been strongly folded into an asymmetrical syncline striking north-west, and modified by subsidiary structural features such as thrust-faults and cross-folds. As a rule the exposures are of highly inclined strata.[4]

Any interpretation of *Lyonothamnus* in the Furnace Creek deposits must be guided by an understanding of the climatic and floristic succession in the western Mohave area during Tertiary time and also by a consideration of the climatic requirements of its assumed nearest living relative. Since the climate over the western Mohave area has been semiarid from at least the beginning of Miocene time, and perhaps since the Middle Oligocene,[5] it is to be expected that the Middle and later Tertiary plant record of the region would be incomplete, because conditions suited to the preservation of leaves are largely unfavorable in arid regions.[6] The woodland and chaparral associations of the western United States, whose related fossil communities contributed largely to the paleobotanical record of southeastern California and adjacent regions in Miocene and Pliocene time, normally occupy lowland sites of deposition where annual precipitation is in excess of about 12 inches. Where rainfall is lower these communities are to be found in moister habitats at higher elevations, and the lowland areas are dominated by the microphyllous sagebrush and desert associations which would contribute but little to accumulating deposits.[7] The absence of any

[4] Written communication, January 10, 1940.

[5] Axelrod, *op. cit.*, p. 57.

[6] R. W. Chaney, "Notes on Two Fossil Hackberries from the Tertiary of the Western United States," *Carnegie Inst. Wash. Pub. No. 349* (1927), pp. 49–56; "The Kucha Flora in Relation to the Physical Conditions in Central Asia during the Later Tertiary," *Geografiska Annaler, in Honor of Sven Hedin* (1935), pp. 75–105.

[7] Although riparian species of ash, cottonwood, sycamore, and willow may occur along streams in these lowland areas, they have no great climatic or stratigraphic significance (Axelrod, "The Stratigraphic Significance of a Southern Element in Later Tertiary Floras of Western America," *Jour. Wash. Acad. Sci.*, Vol. XXVIII [1938], p. 315).

leaf fragments of woodland or chaparral plants in the Furnace Creek playa beds thus suggests that they may have been restricted to cooler and moister canyons adjacent to the Furnace Creek basin. In this connection it should be noted that the occurrence of grasslike or reedlike plants in the playa beds points to an environment similar to that found about the borders of shallow lakes in the arid portions of the western United States. These communities are significant not only in that they would be expected to produce a fossil record comparable to that encountered in the playa sediments along Furnace Creek but because they point also to a climate sufficiently arid to limit tree growth to upland areas. If oaks, piñon pine, juniper, and their regular associates are ever found in the Furnace Creek sediments, they may be expected to have a scanty representation and show signs of transport.

This interpretation not only agrees with the manner of plant distribution in arid regions but is consistent also with the nature of the *Lyonothamnus* specimen, which is an impression of a lateral leaf lobe. The degree of destruction undergone by the leaf before it was buried in the sediments suggests that it may have been transported for a considerable distance to the site of deposition. It may be pointed out that the leaves of the modern Catalina ironwood are coriaceous and tough and do not decay even after lying two or three months on the ground. In arid regions, where leaves would be subjected to strong winds, the drying effects of the sun and transport in torrential waters, erosion, and destruction would be expected to produce a fragment comparable to the Furnace Creek specimen.

From these considerations it may be suggested that the *Lyonothamnus* specimen was derived from a tree occupying a habitat not greatly different from that of its modern occurrence on Santa Catalina Island, where rainfall ranges from 13 to 15 inches yearly.[8] It is not difficult to visualize the Catalina ironwood growing on a rocky and gravelly site on a relatively dry canyon slope or wash adjacent to the Furnace Creek basin, and it is from such an area that the fossil specimen is thought to have been blown or washed into the playa region. Whereas rainfall in the canyon area probably exceeded 12 inches, it is thought to have been several inches lower in the playa

[8] W. L. Jepson, "Silva of California," *Univ. Calif. Mem.*, Vol. II (1910), p. 255.

region where the grasslike or reedlike plants formed communities along the borders of marshes or shallow lakes.

Considering this evidence together with the nature of the Miocene and Pliocene floras of the adjacent region, it is possible to make some suggestions regarding the probable age of the sediments. The Tehachapi flora from the western border of the Mohave Desert indicates that the Middle Tertiary flora of the Mohave region was predominantly a woodland community with associate chaparral and desert scrub species.[9] Species of madrone (*Arbutus*), manzanita (*Arctostaphylos*), buckbrush (*Ceanothus*), hackberry (*Celtis*), mountain mahogany (*Cercocarpus*), fan palm (*Erythea*), piñon pine (*Pinus*), cottonwood (*Populus*), mesquite (*Prosopis*), and various oaks (*Quercus*) ranged widely over the area in early Middle Miocene time, and *Lyonothamnus* occurred here also. Since many of these species are represented in the early Pliocene Esmeralda flora 110 miles north of Death Valley,[10] as well as in the late Miocene and early Pliocene floras farther north,[11] it is to be expected that they ranged also over the Death Valley region during Middle and later Tertiary time. The late Miocene and early Pliocene northward dispersal of this arid woodland flora has been discussed elsewhere.[12] *Lyonothamnus* undoubtedly was a regular member of this flora, for at all its other five fossil occurrences it is associated with typical woodland plants of southern origin.

This arid north Mexican assemblage inhabited the region now desert from Middle Miocene into Lower Pliocene time, when rainfall ranged from a maximum of 25 inches early in the epoch to as low as 10 or 12 inches later in the stage. Since it is believed that *Lyonothamnus* lived in the Furnace Creek area when rainfall ranged from 13 to 15 inches in the hills and canyons to 10 inches or less on the playa flats, it is necessary to determine during which stage of the later

[9] Axelrod, "A Miocene Flora . . . . ," *op. cit.*, pp. 1–129.

[10] Axelrod, "The Pliocene Esmeralda Flora of West Central Nevada," *Jour. Wash. Acad. Sci.*, Vol. XXX (1940), pp. 163–74.

[11] Axelrod, "Late Tertiary Floras of the Great Basin and Border Areas, *Bull. Torrey Bot. Club*, Vol. LXVII (1940), pp. 477–87.

[12] Axelrod, "The Stratigraphic Significance of a Southern Element . . . . ," *op. cit.*, pp. 313–22; "A Miocene Flora . . . . ," *op. cit.*, pp. 49–58.

Tertiary rainfall may have been distributed in this manner. The general climatic trend in the western Mohave area can be tabulated as shown in Table 1. The variation in rainfall indicated by these floras, which has been estimated from the moisture requirements of the modern equivalent species of these assemblages, is considered to be a reflection of habitat differences over the area. If these data are correct, it follows that the upper Furnace Creek formation is prob-

TABLE 1

	Rainfall in Inches
Lower Pliocene Ricardo flora*	10–15
Upper Miocene (estimated from Mint Canyon flora)†	13–20
Middle Miocene Tehachapi flora	15–25

* Axelrod, "A Miocene Flora . . . ," *op. cit.*, pp. 80–81.

† Axelrod, "The Mint Canyon Flora of Southern California: A Preliminary Statement," *Amer. Jour. Sci.*, Vol. CCXXXVIII (1940), pp. 577–85.

ably not older than late Miocene or younger than Lower Pliocene. Although the Ricardo and Furnace Creek areas appear to have had essentially the same amount of rainfall, this does not argue for an age correlation. Since the Death Valley region is situated farther inland than Ricardo, it would be expected to have had a drier climate than existed along the western border of the province during later Tertiary time. The upper Furnace Creek beds thus appear to be older than the Ricardo; and, if they are Upper Miocene, they belong to the late part of this stage.

ORIGIN OF LITTLE NORTH MOUNTAIN IN NORTHERN VIRGINIA[1]

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ABSTRACT

Little North Mountain, the western boundary of the Shenandoah Valley in northern Virginia, is an overturned monocline, with steep southeast dips, bounded on the east by a thrust fault. The formations involved, Upper Ordovician to Lower Devonian, show pronounced variations in thickness, and one or more units are absent at different locations along the mountain. Such discontinuous distribution of sediments, especially the ridge-making Tuscarora sandstone, is thought to have acted as a weakened zone during deformation and determined the locus of the fault and the uplifting and overturning of the formations. Numerous water and wind gaps mark locations where the Tuscarora formation is thin or absent, and the higher altitudes correspond to a thickening of this sandstone.

INTRODUCTION

Location.—Little North Mountain begins at the southwest in a complex of unoriented ridges about 10 miles northwest of Lexington, Virginia. From this locality it extends northeastward for about 150 miles across Rockbridge, Augusta, Rockingham, Shenandoah, and Frederick counties, Virginia (Fig. 1), and Berkeley County, West Virginia, to terminate at the northeast in Bear Pond Mountains, Washington County, Maryland. Approximately 27 miles of the length of the mountain are in West Virginia and Maryland.

The Appalachian Valley and Ridge province, which includes the Little North Mountain area, lies between the northwestern ridges of the Blue Ridge and the eastern escarpment of the Appalachian plateaus. The topography within this province is such that the province can be divided naturally into two distinct parts, which have long been recognized as the Valley of Virginia to the east and the linear parallel ridges, or Valley Ridges, to the west. In northern Virginia the Shenandoah Valley is coextensive with the Valley of Virginia. Little North Mountain in Virginia is the easternmost ridge of the Valley Ridges and, therefore, makes the western boundary of the Shenandoah Valley.

Scope of paper.—The purpose of this paper is to summarize the

[1] Published with the permission of the state geologist of Virginia.

character of the rock units, topography, and structure along Little North Mountain in Frederick and northern Shenandoah counties, Virginia, and to describe in detail the important role of sedimentation in the geologic history of the area. It is thought that the Little North Mountain area was a belt of the sea floor subject to oscillations that resulted in the unequal distribution of sediments during the Oswego, Juniata, Tuscarora, Clinton, and Cayugan epochs.

FIG. 1.—Index map showing extent of Little North Mountain and location of cross sections in Frederick and northern Shenandoah counties, Virginia.

Such discontinuous sedimentation produced a zone of weakness that probably was an important factor in controlling the subsequent structures. Land sculpture and the remarkable influence of the Tuscarora sandstone in producing the varied topography are emphasized.

Previous geologic work.—The first observations on Little North Mountain date as far back as the Virginia Geological Survey of 1835–41, when William B. Rogers included many detailed structure

sections of the mountain in his *Geology of the Virginias*.[2] Many later papers on Virginia geology have mentioned or discussed briefly the area. The first detailed description of the linear mountain belt in northern Virginia was by Giles[3] in 1927. Butts and Edmundson[4] in 1939 gave a more detailed description and interpretation of the area as the result of recent detailed mapping for the Virginia Geological Survey.

Giles[5] interpreted the structure of Little North Mountain to be a

> faulted monocline of overturned beds with steep southeastward dip bounded on the east by closely compressed generally overturned folds affecting the Cambrian and Ordovician formations underlying the Valley, and on the west by a broad syncline developed in Devonian rocks. The faults are overthrusts from the southeast with southeastward inclination. The major overthrust presents a continuous trace on the west side of the mountain. The other faults are distributive from the major overthrust and consequently are discontinuous, merging with the major overthrust at infrequent intervals. Their traces in general lie east of the mountain crest and locally may depart some distance beyond the foot of the mountain.

Butts and Edmundson[6] accept a part of this interpretation of the structure. The overturned beds, with steep southeast dips bounded on the east by faulted overturned Cambrian and Ordovician rocks and on the west by a broad Devonian syncline, are well displayed; but the assumed major overthrust, with a continuous trace on the west side of the mountain, was not identified. The assumed fault is well supported by the stratigraphic interpretations of Giles, but that interpretation is not in agreement with the data obtained by recent detailed stratigraphic studies. Butts and Edmundson differed also from Giles in postulating discontinuous sedimentation, rather than faulting, to account for the marked variations in thickness, and even absence, of one or more beds at different localities along Little North Mountain.

[2] *A Reprint of Annual Reports and Other Papers on the Geology of the Virginias* (New York: D. Appleton & Co., 1884), Pls. 1, 2, 5, 7, and 8.

[3] A. W. Giles, "The Geology of Little North Mountain in Northern Virginia and West Virginia," *Jour. Geol.*, Vol. XXXV (1927), pp. 32–57.

[4] Charles Butts and R. S. Edmundson, "The Geology of Little North Mountain in Northern Virginia," *Va. Geol. Surv. Bull. 51-H* (1939), pp. 164–79.

[5] *Op. cit.*, p. 53. [6] *Op. cit.*, pp. 173–74.

TOPOGRAPHY

General character.—At many places Little North Mountain does not have the form and height[7] of a typical mountain ridge, although it is a structural entity. South of the Potomac River the mountain rises as a linear ridge from an altitude of about 400 feet to 1,640 feet at Roundtop, Berkeley County, West Virginia. It then gradually descends to 1,345 feet at the West Virginia–Virginia boundary. The topographic expression in Frederick and northern Shenandoah counties (Fig. 2) is discussed below under "Detailed Features."

The ridge is absent in southern Shenandoah County but reappears boldly to the south in northern Rockingham County. It is lacking in southwest Rockingham and northern Augusta counties. In southern Augusta County it is expressed topographically as a discontinuous ridge which extends southwestward to the complex of ridges near Goshen Pass, in northwestern Rockbridge County.

Detailed features.—The summit of the mountain[8] at the Virginia–West Virginia boundary is 1,345 feet. From this point the crest gradually descends to 900 feet and then abruptly drops to 690 feet in the Green Springs Gap. The conical hill, with an altitude of 800 feet, just south of Green Springs Run is followed for 1 mile to the southwest by low, hummocky topography.

Beginning at Cattail Run, the longitudinal profile gradually steepens to the crest of Babbs Mountain and then descends abruptly to Babbs Run. Babbs Mountain, with a maximum relief of 300 feet when viewed from a distance, appears to be an isolated knob on the adjacent valley floor. The next sector of the mountain, Flint Ridge, extends from Babbs Run southwest to Pearls Run, a distance of about $3\frac{1}{2}$ miles. The crest of this narrow ridge rises, on the average, about 300 feet above the surrounding area and, except for the wind gap in which the village of Nain is located, slopes gradually southwestward. From Pearls Run southwest to the northeast base of Round Hill (Fig. 3, *A*) there is a small discontinuous ridge, which at no place is more than 50 feet above the surrounding valleys.

Round Hill rises steeply to an altitude of 1,397 feet and then

[7] The general features are shown on the topographic maps of the Gerrardstown, Winchester, and Middletown quadrangles.

[8] All references to geographic elements and altitudes are to those in Fig. 2.

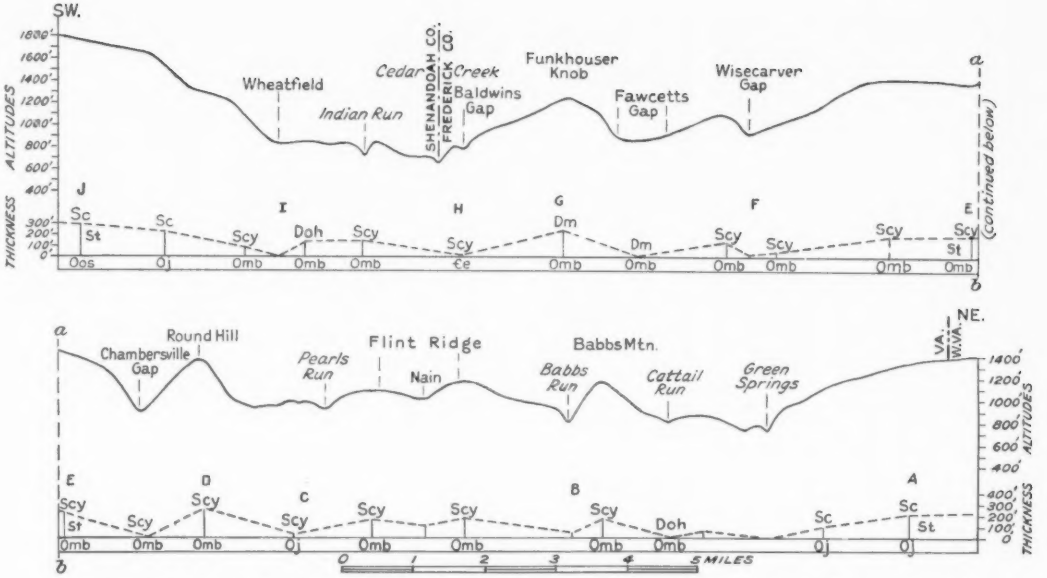


FIG. 2.—Relation of the profile of Little North Mountain in northern Virginia to the distribution and thickness of the underlying Tuscarora sandstone (*St*). Contiguous formations include Martinsburg shale (*Omb*), Oswego sandstone (*Oos*), Juniata shale (*Oj*), Clinton shale and sandstone (*Sc*), Cayugan shale and sandstone (*Scy*), Helderberg limestone and Oriskany sandstone (*Doh*), and Marcellus shale (*Dm*). *A–J* show locations of geologic sections in Fig. 6.

sharply descends to Gap Run, just west of the village of Chambersville along U.S. Route 50. This hill, with a maximum relief of 550 feet, is an outstanding landmark in northern Virginia.

Beginning just south of Gap Run, the mountain rises as a bold ridge, trending S. 35° W., and maintains an even crest (Fig. 3, *B*) for a distance of about 2½ miles and then gradually diminishes in height to Wisecarver Gap. The maximum altitude along this profile is 1,474 feet, and it averages about 1,400 feet. Wisecarver Gap (Fig. 4, *A*), with an altitude of 900 feet, is bounded on the southwest by an elongate hill that rises steeply from the gap and gradually descends to Fawcetts Gap, 1½ miles to the southwest. The altitude of the summit of the hill is 1,106 feet. Fawcetts Gap, with an average altitude of 820 feet, is a broad flat area along the trend of the mountain that connects laterally with the adjacent valleys.

The ridge between Fawcetts and Baldwins gaps, a distance of 2½ miles, culminates in Funkhouser Knob (Fig. 4, *B*) at an altitude of 1,282 feet. The part of the mountain between Baldwins Gap and the village of Wheatfield, Shenandoah County, includes the narrow valley of Cedar Creek and its tributary, Indian Run. The bottom of Cedar Creek gorge, with an altitude of about 660 feet, is the lowest point along Little North Mountain in Frederick and northern Shenandoah counties. The topography along this section of the mountain consists of a series of disconnected low elongate hills, averaging in height about 100 feet above the surrounding country.

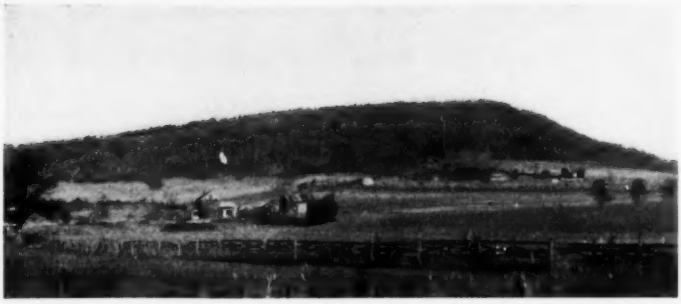
Beginning just south of Wheatfield, Little North Mountain becomes a conspicuous ridge (Fig. 4, *C*) which maintains an even crest at an average altitude of about 2,000 feet for several miles in Shenandoah County, south of the area discussed here.

STRATIGRAPHY

The Little North Mountain area includes twenty-one formations (Table 1) that range from Middle Cambrian to Upper Devonian, inclusive, and have an aggregate thickness of about 17,000 feet.

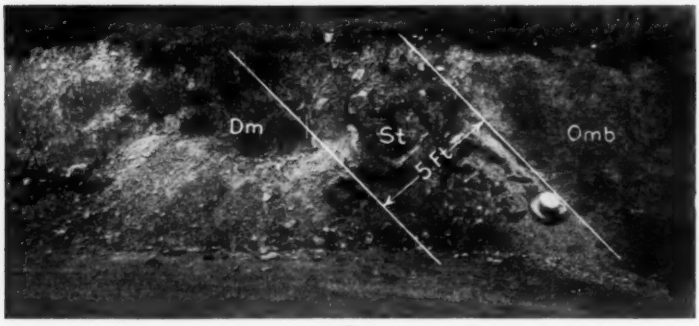
In a recent report[9] the lithology of each unit was described and a few diagnostic guide fossils were given, but in this paper the discussion is limited to those formations that are included in the mountain

[9] Butts and Edmundson, *op. cit.*, pp. 166–73.



A

B



C

FIG. 3.—*A*, Round Hill on Tuscarora sandstone. The sandstone is approximately 150 feet thick at the summit of the hill and thins to about 50 feet within the limits of the photograph. Looking west from U.S. Route 50, Frederick County, Virginia.

B, Even crest of Little North Mountain between Round Hill and Wisecarver Gap. The crest is on Tuscarora sandstone, which forms a part of the overturned monocline with steep southeastward dips. Looking west from a point 1 mile northeast of Opequon, Frederick County, Virginia.

C, Silurian system in Fawcetts Gap, Frederick County, Virginia. The Silurian here is only 5 feet thick. It consists of the Tuscarora sandstone. The beds dip southeast. *Omb*, Martinsburg shale; *St*, Tuscarora sandstone; *Dm*, Marcellus shale. Photograph by Arthur Bevan.

A

B

C

FIG. 4.—*A*, Wisecarver Gap, Frederick County, Virginia. The Tuscarora sandstone is approximately 30 feet thick in the gap, 150 feet on the left, and 50 feet on the right. Great North Mountain in the distance.

B, Funkhouser Knob southwest of Fawcetts Gap, Frederick County, Virginia. The knob is underlain by a lens of Tuscarora sandstone about 150 feet thick. Looking southwest along trend of mountain.

C, Little North Mountain southwest of Wheatfield, northern Shenandoah County, Virginia. The ridge here has an average altitude of 2,000 feet.

ridge and which probably had an important part in its geologic history. These formations include sediments of the Oswego to Oriskany epochs, inclusive, having a total maximum thickness of not more than 1,600 feet (see Table 1).

TABLE 1

FORMATIONS OF THE WESTERN PART OF THE VALLEY IN NORTHERN VIRGINIA

Period or System	Formation	Maximum Thickness (Feet)
Devonian	Hampshire formation	2,000
	Chemung formation	2,000
	Brallier shale	1,500
	Hamilton formation	1,000
	Marcellus shale	500
	Onondaga shale	200
	Oriskany sandstone*	100
	Helderberg limestone*	300
Silurian	Cayugan group*	400
	Clinton formation*	200
	Tuscarora sandstone*	300
Ordovician	Juniata formation*	100
	Oswego sandstone*	200
	Martinsburg shale	2,000
	Chambersburg limestone	600
	Lenoir limestone	50
	Mosheim limestone	50
	Beekmantown limestone	2,500
	Chapultepec limestone	600
Cambrian	Conococheague limestone	2,000
	Elbrook limestone	2,000

* Formations included in the Little North Mountain ridge.

UPPER ORDOVICIAN

Oswego sandstone.—The Oswego sandstone lies between the *Orthorhynchula* zone at the top of the Martinsburg formation and the Juniata formation. It crops out only in the northern part of Shenandoah County, southwest of Wheatfield, where it consists of thick-bedded, gray, iron-stained sandstone and conglomerate. A few thin shale partings are found locally. The thickness ranges from a few feet southwest of Wheatfield to about 200 feet at Pontzer Gap in Shenandoah County.

Juniata formation.—The Juniata formation, which lies conformably on the Oswego sandstone, consists of red shale and mudrock and beds of brown to red sandstone. The distinctive red color makes it one of the easiest formations to recognize. Along Little North Mountain the formation occurs as discontinuous lenses ranging from a few feet to a possible maximum of about 100 feet in thickness.

SILURIAN

Tuscarora sandstone.—The Tuscarora sandstone is composed of light-gray to white pure quartz sand in a siliceous matrix. It is a compact vitreous rock commonly called a "quartzite." In Frederick and northern Shenandoah counties the Tuscarora occurs as lenses that range in thickness from a few feet to approximately 300 feet.

Clinton formation.—The Clinton is well displayed north of Green Springs, Frederick County, and south of Wheatfield, Shenandoah County; but it is absent in every exposed section between these points through a distance of about 20 miles. It succeeds the Tuscarora and extends upward, apparently, to the Bloomsburg member of the Cayugan group. The Clinton is composed of yellowish shale and thin ferruginous sandstone, which have an aggregate thickness of about 200 feet.

So far as known, the Keefer sandstone member at the top of the Clinton and the McKenzie formation (Lower Cayugan) are not present along Little North Mountain.

Cayugan group.—The Cayugan group is complete and well displayed at the northeastern end of Great North Mountain, along U.S. Route 50 about 4 miles west of Chambersville; but along Little North Mountain only the Bloomsburg, Wills Creek, and Tonoloway divisions have been identified. The Bloomsburg formation is composed predominantly of red shale and mudrock and thin red sandstone. At a few localities in Frederick County a massive sandstone, about 10 feet thick, is a persistent marker of the upper part of the Bloomsburg. The red shales and sandstones are succeeded by yellowish shale of the Wills Creek formation. The aggregate thickness of the two formations must be about 300 feet. In Baldwins Gap about 100 feet of platy limestone containing Tonoloway fossils overlies the Wills Creek formation. The lack of adequate exposures

makes it impossible to determine the extent of the Tonoloway along the mountain, but in many sections examined the Wills Creek is succeeded by the Helderberg limestone.

LOWER DEVONIAN

Helderberg limestone.—The lowermost Devonian is represented by the Keyser and New Scotland limestones. A few sections show the Keyser to be, in part, a thick-bedded nodular limestone with a maximum thickness of about 200 feet, which is succeeded by a zone characterized by the development of white prismatic chert fragments containing characteristic New Scotland fossils. Although the New Scotland is rarely exposed, the distribution of the chert suggests a thickness of 50–100 feet.

Oriskany sandstone.—The Oriskany sandstone is not a persistent bed. In some sections it is definitely absent, whereas it is a few feet thick at other localities. The best development in this belt is about 1 mile northeast of Round Hill, Frederick County, where the formation is not less than 100 feet thick. The maximum thickness in Frederick County is about 200 feet. The Oriskany is a thick-bedded, light-colored calcareous sandstone and conglomerate.

GEOLOGIC HISTORY

SEDIMENTATION

At the beginning of the Paleozoic era the site of Little North Mountain was a very small part of a broad lowland occupied by igneous and metamorphic rocks, probably similar to some rocks now exposed in the Piedmont province. The oldest sedimentary rocks deposited in the Appalachian geosyncline consist chiefly of quartz sand and fragments of feldspar, which were derived from the pre-Cambrian rocks to the east. These clastic sediments (Lower Cambrian), although deeply buried to the west, crop out along the west slope of the Blue Ridge and to the east in the Piedmont province. The overlying rocks, deposited during late Cambrian and the pre-Martinsburg epochs of the Ordovician period, are chiefly limestone and dolomite, which crop out extensively in the Shenandoah Valley. The Martinsburg epoch is well represented by several thousand feet of shale and fine-grained sandstone with some limestone beds along

the southeast base of Little North Mountain and to the east in the Massanutten syncline near the middle of the Shenandoah Valley. The sediments of the Oswego epoch, identified only in northern Shenandoah County, are coarse quartz and feldspar fragments. The fact that the Oswego has its maximum development in Pennsylvania and is not known east of Little North Mountain suggests that the source of the sediments was to the north. The uppermost Ordovician Juniata shale occurs along Little North Mountain only as thin disconnected bodies.

The basal Silurian (Tuscarora) sediments, according to Butts,[10] covered an area from central Pennsylvania to Tennessee. They probably originated in a vast sand flat near sea-level, which was submerged intermittently by the tide, by which at times the sediment was distributed. At other times they were possibly swept by storms which blew away the finer sediment. The residue, like ordinary beach sand, was transported seaward and formed the beds of the resistant, ridge-making Tuscarora sandstone. During the Clinton epoch that part of the Appalachian trough northwest of the mountain was submerged. As no Clinton is certainly known to be present to the southeast, it is possible that the two detached belts of Clinton shale and sandstone along Little North Mountain mark approximately the southeast shore line of the Clinton Sea.

The sediments of the Cayugan, Helderberg, and Oriskany epochs are widespread in the Appalachian Valley, but the sedimentary records are incomplete in northern Virginia. The different types of sediment, the variations in their thickness, and the absence of one or more formations in relatively short distances suggest that the deposits of these epochs along the Little North Mountain belt were probably laid down in a shallow restricted sea with considerable shifting of the areas of deposition.

The widespread middle and late Devonian seas received fairly continuous and very thick deposits of silts, muds, and sands in northern Virginia. Post-Devonian sediments have been eroded away in Little North Mountain.

[10] Charles Butts, "The Geology of the Appalachian Valley in Virginia," *Va. Geol. Surv. Bull. 52* (in press).

STRUCTURE

General statement.—The structure of Little North Mountain is an overturned monocline with steep southeast dips, bounded on the east by the Little North Mountain fault. This fault (Fig. 6), whose trace now lies east of the ridge, is thought to have extended above the present summit of the mountain in a tightly compressed overturned fold. Subsequently much of the thrust block was removed by erosion, and at the present time only the overturned northwest limb of the anticline is preserved. The hypothetical progressive stages in the development of the mountain are given in Figure 5.

It is also believed that the structure was largely controlled by Paleozoic sedimentation. When compressional forces became active late in the Paleozoic era, the area of Little North Mountain, with its initial sedimentary irregularities, acted as a weakened zone which determined the locus of the fault and the uplifting and overturning of the formations.

Detailed features.—The following structural discussion is limited to the Tuscarora sandstone, or principal ridge-forming formation, the immediately adjacent Juniata, Oswego, and Martinsburg formations on the east, and the Clinton, Cayugan, Helderberg, and Oriskany formations west of the crest of the ridge. The other formations shown in the geologic sections (Fig. 6) are fairly constant in their distribution and are not considered as determining factors in the topographic expression of Little North Mountain.

A complete stratigraphic sequence of the mountain in Frederick County is probably represented in Figure 6, *A*, whereas Figure 6, *J*, shows normal conditions in northern Shenandoah County. So far as could be determined, most of the formations thicken to the southwest, and in that direction a new member, Oswego sandstone, is introduced. Along Section 6, *A*, just south of the Virginia–West Virginia line, all of the formations strike N. 20° E. and are overturned with steep southeastward dips.

The Silurian formations diminish in thickness from the state line to a point just north of the gap of Green Springs Run, where the Martinsburg shale is in contact with Helderberg limestone. Since the strike of the Martinsburg shale does not conform to the structure

of the younger formations, it is reasonably certain that a minor fault, in conjunction with pronounced thinning, must be postulated

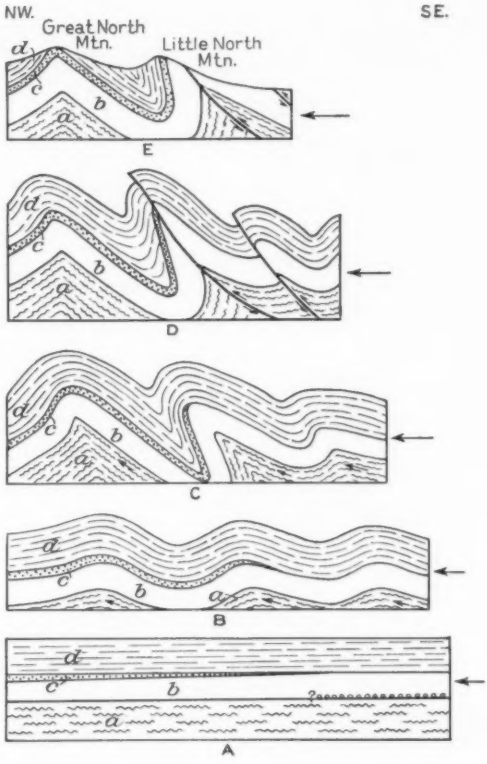


FIG. 5.—Hypothetical progressive stages in the development of Little North Mountain in northern Virginia. *A*, undisturbed section showing (*a*) crystalline basement rocks, (*b*) Cambrian to Upper Ordovician limestones and dolomites, (*c*) Upper Ordovician and Silurian sandstones and shales, and (*d*) Devonian shales and sandstones; *B*, incipient folding; *C*, advanced folding; *D*, thrust faulting; and *E*, present surface.

at this locality. The extent of this fault is not definitely known; but 2 miles to the southwest, on the northeastern slope of Babbs Mountain, the sequence west of the Tuscarora sandstone is normal.

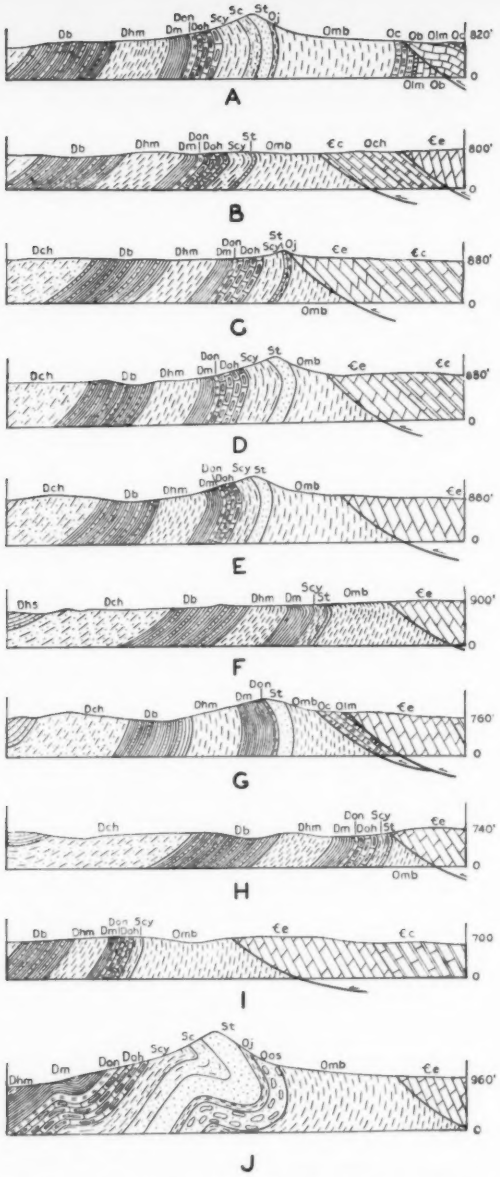


FIG. 6.—Geologic cross sections of Little North Mountain in northern Virginia (see Figs. 1 and 2). *A*, half a mile south of state line; *B*, through Cedar Grove; *C*, 2 miles north of Chambersville; *D*, Round Hill; *E*, 1 mile south of Chambersville; *F*, Wisecarver Gap; *G*, Funkhouser Knob; *H*, Baldwins Gap; *I*, Wheatfield; and *J*, Pontzer Gap. *Ce*, Elbrook limestone; *Cc*, Conococheague limestone; *Och*, Chapultepec limestone; *Ob*, Beekmantown limestone and dolomite; *Olm*, Mosheim and Lenoir limestones; *Oc*, Chambersburg limestone; *Omb*, Martinsburg shale; *Oos*, Oswego sandstone; *Oj*, Juniata formation; *St*, Tuscarora sandstone; *Sc*, Clinton formation; *Scy*, Cayugan formations; *Doh*, Helderberg and Oriskany formations; *Don*, Onondaga shale; *Dm*, Marcellus shale; *Dhm*, Hamilton formation; *Db*, Brallier shale; *Dch*, Chemung formation; and *Dhs*, Hampshire formation.

At the village of Cedar Grove along Babbs Run (Fig. 6, *B*), the Tuscarora sandstone is about 50 feet thick. From this point it can be traced as a lens increasing in thickness to about 150 feet on Babbs Mountain and Flint Ridge. The absence of the Juniata east of the Tuscarora and the Clinton to the west is believed to represent interruptions in sedimentation, since the Clinton does not appear again along the mountain for about 20 miles to the southwest and the Juniata is present as very thin lenses in only a few isolated localities.

The Tuscarora sandstone ranges in thickness from about 150 feet just south of Nain to only a few feet at the southwestern end of Flint Ridge. Favorable bedrock exposures at this locality fail to afford any indications of discordant dips, drag folding, or other criteria of faulting. Midway between Flint Ridge and Round Hill the Tuscarora has a thin body of Juniata along its eastern boundary. In the same section the major fault east of the mountain has cut out the Martinsburg shale and has brought Elbrook limestone in contact with the Juniata. It should be noted that the Tuscarora here is very thin in comparison with its thickness along Flint Ridge, yet the presence of the Juniata indicates an orderly sequence.

Along the northeast slope of Round Hill (Fig. 3, *A*), the Tuscarora begins as a thin formation, rapidly increases to a maximum thickness of about 250 feet (Fig. 6, *D*) near the summit, and then gradually thins out just west of Chambersville. The sequence of overturned beds shown in Figure 6, *D*, is Martinsburg, Tuscarora, Cayugan, and Helderberg; whereas in the gap, 1 mile to the southwest, the Martinsburg shale is in contact with Cayugan shales. Giles[11] interpreted the geology here as Clinton thrust upon Helderberg, but fossils show that the supposed Clinton is Cayugan and that the sequence therefore is conformable for the Upper Silurian and Lower Devonian beds. To interpret the display of the Tuscarora sandstone in Round Hill as a depositional lens is speculative; yet the attitude of the enclosing rocks plainly suggests such an origin.

The Tuscarora reappears as a thin bed just south of the gap west of Chambersville and rapidly thickens to about 200 feet. This average thickness (Fig. 6, *E*), which produces an even crest line, is main-

[11] *Op. cit.*, sec. 9, p. 55.

tained for 3 miles to the southwest. The formation then gradually thins to 30 feet in Wisecarver Gap (Fig. 6, *F*). The sequence in Wisecarver Gap (Fig. 4, *A*) differs from that represented by preceding sections in that the Helderberg and Onondaga formations are absent. It suggests that discontinuous sedimentation locally extended through a part of early Devonian time. The prominent knob just south of Wisecarver Gap (Fig. 4, *A*) is underlain by at least 150 feet of Tuscarora sandstone. This local thickening of the sandstone resembles the occurrences at Round Hill and Babbs Mountain. From this point the Tuscarora thins gradually southwestward to about 5 feet in Fawcetts Gap (Fig. 3, *C*), where it is flanked on the east by Martinsburg shale and on the west by Marcellus shale. The absence of the Cayugan, Helderberg, and Onondaga formations above the Tuscarora sandstone represents the most extensive hiatus along Little North Mountain, although the total thickness of these missing beds at other near-by localities does not exceed 500–600 feet.

Beginning on the northeast slope of Funkhouser Knob (Fig. 4, *B*), the sandstone increases rapidly in thickness to about 250 feet at the summit (Fig. 6, *G*). The overturned beds, with steep southeast dips, are similar to those in the Fawcetts Gap section, except for the thick Tuscarora. Just southwest of the summit of Funkhouser Knob the Onondaga, Helderberg, and Cayugan formations wedge out between the Tuscarora sandstone and the Marcellus shale.

At Baldwins Gap (Fig. 6, *H*) the Elbrook limestone is thrust upon 20 feet of Tuscarora sandstone. West of the Tuscarora sandstone the overturned sequence includes Cayugan, Helderberg, and Onondaga; whereas it was interpreted by Giles[12] as Cayugan thrust upon Romney shale several hundred feet above the base of that formation. Between Baldwins Gap and Wheatfield, Shenandoah County, there are two disconnected lenticular bodies of Tuscarora, with a maximum diameter of about 150 feet. Along the lens at the north is a thin body of Juniata shale. The southwestern body thins out about 1,000 feet northeast of Wheatfield. At this locality (Fig. 6, *I*) the Martinsburg is followed by Cayugan, Helderberg, and Onondaga.

[12] *Ibid.*, sec. 12, p. 55.

Southwest of Wheatfield and extending to the southwestern boundary of the area under consideration (Fig. 4, *C*), all the formations with their maximum thickness are introduced. In the vicinity of the area shown in Figure 6, *J*, the Tuscarora sandstone is about 300 feet thick. To the east the normal sequence includes about 100 feet of Juniata, about 200 feet of Oswego sandstone and conglomerate, and a wide expanse of Martinsburg shale in fault contact with Elbrook limestone along its eastern boundary. To the west the widths of the various outcrop belts suggest 200 feet of Clinton, 400 feet of Cayugan, 200–300 feet of Helderberg and Oriskany, and 200 feet of Onondaga.

LAND SCULPTURE

The sculpture of Little North Mountain dates back to the close of the Paleozoic era, but the present topographic features in northern Virginia do not afford evidences of the subsequent complicated erosional history. The low relief of Little North Mountain is interesting when it is compared with the neighboring mountains to the west. For example, Great North Mountain rises as a bold ridge in northern Frederick County and continues to the southwest along, and parallel to, the part of Little North Mountain that exhibits an unusual amount of dissection. The above differences in topography are believed to have been largely controlled by sedimentation and to a less degree by the structure. Structurally, Great North Mountain is a broad anticline modified by subsidiary flexures along the flanks, whereas Little North Mountain is a monocline of steeply dipping beds (Fig. 6). The anticline, with relatively resistant beds repeated, has been more resistant to erosion than the faulted monocline. Of still greater importance is the fact that all studied sections show marked stratigraphic regularity along Great North Mountain, whereas varied discontinuities in Frederick County are the rule in Little North Mountain.

Sedimentation, especially in the Tuscarora epoch, has been the important factor in determining the topography of Little North Mountain (see Fig. 2). Along the mountain the water and wind gaps, exclusive of Cedar Creek, which probably had its course well

established on the Upland peneplain before the succeeding uplifts, occur at places where the Tuscarora sandstone is thin or absent, and the higher altitudes correspond to a thickening of this sandstone. The role of the Tuscarora sandstone as a ridge-maker is well shown by most of the Valley Ridges, but Little North Mountain shows a more exacting control. In northern Shenandoah County the mountain is conspicuous (Fig. 4, *C*), with altitudes of more than 2,000 feet; but that is due to the fact that the ridge-making Oswego and Clinton formations, in addition to the Tuscarora, are well developed. The role of the Tuscarora sandstone in the history of Little North Mountain is more readily ascertained in areas where the enclosing rocks are relatively weak shales and sandstones. In this environment the tributary streams cross the mountain where the Tuscarora sandstone is thin or absent, and they do not necessarily coincide with any previous location on an old erosion surface (see Fig. 2).

SUMMARY

The data concerning the form, structure, and geologic history of Little North Mountain can be summarized as follows: The ridge in Frederick and northern Shenandoah counties contains many water and wind gaps. The formations, including Oswego, Juniata, Tuscarora, Clinton, Cayugan, Helderberg, and Oriskany, have pronounced variations in thickness, and one or more units are absent at different localities. The water and wind gaps occur at places where the Tuscarora sandstone is thin or absent and the higher altitudes correspond to a thickening of this sandstone. The structure is an overturned monocline, with steep southeast dips, bounded on the east by a thrust fault which may crop out as much as half a mile east of the base of the mountain.

The discontinuous distribution of the sediments suggest that the Little North Mountain area was an unstable belt of the sea floor subject to many oscillations during late Ordovician, Silurian, and early Devonian time. It is thought that, when compressional forces became active late in the Paleozoic era, the area of Little North Mountain, with its initial sedimentary irregularities, acted as a weakened zone which determined the locus of the fault and the up-

lifting and overturning of the formations. All the stratigraphic variations along and immediately west of the mountain crest, except for possible minor displacements accompanying intense local folding, are due to irregularities of sedimentation rather than to faulting.

ACKNOWLEDGMENTS.—The writer is indebted to Dr. Charles Butts for the interpretation of many of the complex stratigraphic problems in northern Virginia, studied jointly in the field during the summers of 1935–39. The joint paper previously mentioned, although stressing different elements of the geology, has served as a basis for the present interpretation. Additional aid has been afforded by having access to the manuscript by Dr. Butts on "The Geology of the Appalachian Valley in Virginia."[13]

[13] *Va. Geol. Surv. Bull. 52* (in press).

REVIEWS

Foraminifera, Their Classification and Economic Use. By JOSEPH A. CUSHMAN. Cambridge, Mass.: Harvard University Press, 1940. Pp. viii+480. 86 illustrations. $6.00.

This is the third revised and enlarged edition of the book which has come to be rather universally regarded as the standard guide to the Foraminifera. Less than twenty-five years ago these microscopic organisms were of interest to only a few specialists, although a small group of Italian scientists in Venice and vicinity started to examine and describe them in an amateurish way as early as 1735. At present, largely because of their having been made available for economic work in connection with prospecting and drilling for petroleum, there are hundreds of workers all over the world devoting much of their time to the investigation of the Foraminifera from the utilitarian point of view. The author of the present guidebook may well be pardoned a considerable amount of pride in the outstanding role which he himself has played in this modern development of, and growth of interest in, both the scientific and economic phases of studies of these microscopic fossils.

The early editions of the manual were published privately by Dr. Cushman—the first in 1928, the second in 1933. The latter edition included a second volume made up of a key to the families and genera, and forty full-page plates. The third edition has been issued as a single volume with the plates and text under a single cover. Although the third edition contains a number of genera not appearing in the second, there have been, on the whole, remarkably few changes. Eight new Key plates have been added, but Plates I–XL are the same as those in the second edition. Thus, the genera newly illustrated have been appended rather than placed in their correct positions so far as families are concerned. The text treatment of the phenomenon of trimorphism is still, as in the second edition, widely separated from the discussion of dimorphism, and, although Meyers' work is listed in the Bibliography, Cushman makes no statement of Meyers' concepts concerning the life-cycle of the Foraminifera in his text. The Bibliography, which is one of the most valuable features of the book, has now been extended to sixty pages; and, although it is, of course, incomplete, it nevertheless apparently includes most of

the really significant works on the Foraminifera arranged according to subjects, geological age, and geographic distribution. A number of papers issued in 1939 are included.

There is not much deviation in the third edition from the classification adopted in the earlier issues of the manual. Altogether approximately five hundred genera are recognized as valid, not including a few genera which have been omitted because of the difficulty of placing them correctly. The genera recognized are distributed among forty-nine families, of which the Discocyclinidae and the Miogypsinidae are added. These new families are proposed by T. Wayland Vaughan and W. Storrs Cole in this volume, and the sections describing them appear under the names of these specialists.

The Harvard University Press has dressed up this important publication in a handsome and durable cover and has done a good job of reproduction and printing. The expanded treatment under a single cover marks a notable improvement, despite the fact that the text changes made probably are not so great as might have been expected. The insertion of the Key plates in the single volume makes for the anomalous situation of two similarly numbered series of plates. There are also seven illustrations lacking sequential number designations, which might be regarded as text figures. These, of course, are minor editorial imperfections in a publication which will continue to be regarded as indispensable to all micropaleontologists and to most stratigraphical paleontologists.

CAREY CRONEIS

Geological Expedition to the Lesser Sunda Islands under the Leadership of H. A. Brouwer, Vol. I. Amsterdam: Noord-Hollandsche Uitgevers Maatschappij, 1940; New York: Nordeman Pub. Co., 1940. Pp. 348. Illustrated. $8.40.

This is the first of a contemplated four-volume series whose last three volumes may well be considerably delayed on account of the European war. Professor Brouwer of the University of Amsterdam, the director of the project, for a number of years carried on field work with his students in the Betic Cordillera. When these studies were interrupted in 1936 by the Spanish Civil War, Professor Brouwer proposed a colonial expedition in which students could participate with a view to finishing the necessary field work for their Doctors' theses. From earlier work Dr. Brouwer was well aware that the Netherlands East Indies include large territories which are scarcely explored and that geologic problems of the

utmost importance in that area await closer study. He decided, therefore, to conduct geological explorations in the two rows of the Lesser Sunda Islands north of Australia. Research in this region promised to contribute much in the understanding of the growth of folded mountain ranges as well as in the relationship between tectonic and volcanic evolution. Almost no geological work had been done in some of the islands of the northern row; and Timor, which has become structurally and paleontologically one of the most intriguing of all geological areas, was still insufficiently known.

The organization of the work was facilitated by the fact that Mr. F. A. H. W. de Marez Oyens and Professor Brouwer had already worked in the region in 1910–12, both as members of the well-known Molengraaff expedition and independently. Moreover, Professor Wanner of Bonn, long noted for his studies on the geology and paleontology of the western part of Timor, was able to give valuable advice during the preparation of the expedition. The expedition was financed through the liberal support of a number of industrial, commercial, and scientific bodies connected with the colonies, as well as by the Netherlands colonial government itself, and the municipal authorities of Amsterdam and the curators of the university of that city. The work was carried on during the dry season of June–November, 1937, and the work of the party was facilitated by the use of a number of unpublished topographic maps as well as of native topographers who were put at their disposal. The field work included exploration in the islands of Wetar, Lirang, Lomblen, Batoe, Tara, Adonara, Solor, Eastern Flores, and West or Netherlands Timor.

The results of the individual investigations of the members of the expedition, together with a few paleontological and petrological contributions by other collaborators, will be published in the first three volumes, whereas Volume IV will contain papers co-ordinating the different contributions and will give such general conclusions as doubtless will become apparent as a result of the work.

The present volume contains the following sections: D. Tappenbeck, "Geologie des Mollogebirges und Einiger Benachbarter Gebiete (Niederländisch Timor)," pp. 1–106; A. L. Simons, "Geological Investigations in N.E. Netherlands Timor," pp. 107–214; J. Wanner, "Neue Blastoideen aus dem Perm von Timor (mit einem Beitrag zur Systematik der Blastoideen)," pp. 215–82; F. A. H. W. de Marez Oyens, "Neue Permische Krinoiden mit Angaben über deren Fundstellen im Basleo Gebeit (Niederländisch Timor)," pp. 283–348.

Dr. Tappenbeck's study chiefly concerns the stratigraphy and structure of the Mollo Range and adjacent areas in Netherlands Timor. The stratigraphic sequence includes, in ascending order, (1) crystalline schists of uncertainly known age but which are probably pre-Carboniferous, (2) Permian beds comprising both crinoidal limestones and a volcanic series, (3) Triassic rocks belonging to three different groups, (4) Cretaceous and probably Jurassic strata, and (5) both older and younger Tertiary sequences. The strata of the Mollo Mountains are complexly folded and faulted; and the structure is, as yet, incompletely understood. Stratigraphic tables for the Permian and Mesozoic in the Mollo Mountains, as well as for the Tertiary beds in the Mollo, Lalan-Assoe, and Booi Mountains, are useful additions to the text. Altogether eight plates, some of them made up of a number of figures, and a generalized geological map accompany the paper. Unfortunately, the plates and map for Tappenbeck's contribution are misplaced, having been transposed into the position which should have been occupied by the illustrations for Dr. Simons' article.

Dr. Simons' paper, which is the only one of the four contributions written in English, deals with the geology of northeastern Netherlands Timor, chiefly in the subdivision Beloe, and especially the districts Djeniloe and Lidak.

Essentially the same general rock sequence as described by Dr. Tappenbeck occurs in the regions investigated by Dr. Simons. The Permian deposits are scattered, and the dimensions of the outcrops are small. As a consequence, the complete sequence of Permian beds is not known, nor has it been possible to describe a definite stratigraphic succession of the strata. As is true in most of the Timor areas, both Portuguese and Netherlandish, the Triassic is made up of three different facies—the Kekneno series, Sonnebait series, and the Fatoe complex. These three differ widely from one another not only as to sedimentological composition but also as to organic content. Only a few Jurassic strata can be definitely identified, but both the Cretaceous strata and the Tertiary rocks are fairly extensively represented.

The petrology of the definitely determinable Permian and Mesozoic igneous rocks has received much previous attention, and the ages assigned are considered reasonably certain. There are also a number of "volcanic conglomerates," considered to be Tertiary age; but actually, up to the present the various age interrelations of all these igneous rocks have not been satisfactorily determined. Simons' paper, like that of Dr.

Tappenbeck, has been illustrated by reproducing what apparently are rough field sketches and sections. Many of these could have been redrawn with considerable profit. The photographic reproductions in both articles, moreover, leave a great deal to be desired.

Professor Wanner's contribution on the Permian blastoids of Timor is a significant one and probably will receive more world-wide attention than any of the other papers. This is true not only because of the almost unbelievably bizarre character of this youngest of the blastoidean faunas but because Dr. Wanner has included a general philosophical discussion of the classification of the blastoids. In this work he resurrects Jaekel's two orders Fissiculata and Spiraculata, and he includes under each of these two orders a number of families, most of which he himself has erected. Professor Wanner's paper is well illustrated by three plates and a number of text figures.

Dr. de Marez Oyens' contribution on the Permian crinoids from the Basleo beds, although perhaps less significant from the standpoint of classification, is another important contribution to our knowledge of the peculiar echinoderm fauna of the late Paleozoic Timor seas. Some of the specimens illustrated on the four excellent plates accompanying his paper are sufficiently bizarre almost to tax the credulity of the paleontologist not already familiar with the extraordinary Timor material which has steadily been coming to light since Professor Wanner's first startling publications on the subject in 1916.

The first volume in Professor Brouwer's series contains papers of sufficient value to geologists everywhere so that the next three volumes will be awaited with considerable anticipation; and the expedition organizer and leader deserves, and undoubtedly will receive, great credit for carrying to a successful conclusion this important undertaking. It will be hoped, however, that subsequent contributions will be more standardized, especially as to the quality and general perfection of the editorial work, and that the final volume will actually contain, as promised, sufficiently broad generalizations to permit the geologist who is pretty much puzzled by the details of obscure Sunda Islands localities to gain a more clarified picture of the general geology of the entire archipelago. It also would undoubtedly increase the use of this important series if, somehow, the almost exorbitant price of the individual volumes could be rather substantially reduced.

CAREY CRONEIS

Kemp's Handbook of Rocks. By F. F. GROUT. New York: D. Van Nostrand Co., 1940. Pp. viii+300; figs. 96. $3.00.

The sixth edition of this well-known work has been so completely re-written as to constitute a new book. Nevertheless, the original objectives and spirit of the earlier editions are retained. The newer handbook, like its predecessors, is an elementary guide, largely descriptive, to the study of rocks. It is written for beginning students and others who do not have access to or skill needed to use the petrographic microscope.

As in the earlier editions, the igneous rocks are described first, followed by the sedimentary and metamorphic rocks. Each section is preceded by an introductory chapter on the textures and structures of each group and the relation of these factors to the occurrence and origin of the class. It is impossible, even in a descriptive work, to divorce the problems of origin from the descriptions and classification. The chapters on the various families of rocks systematically cover general description, mineralogic composition, varieties, geologic occurrence, uses, alteration and metamorphism, and distribution. A general summary, including some remarks on genesis, follows the treatment of each of the three principal classes of rocks. The book contains, also, a chapter on calculations in rock study and a chapter on rock descriptions with numerous examples. An appendix of rock names not in the Index, with a single-word definition or synonym, takes the place of the more extended Glossary of the earlier editions. A short foreword in appreciation of James Furman Kemp was written by Professor C. P. Berkey.

Most striking differences between this and earlier editions, other than format, are the introduction of a section on logging wells, the section on rock descriptions, omission of the many tables on chemical composition of the older editions, omission of the tables used in calculation of the mineral composition from chemical analyses, and a great condensation of the Glossary.

This handbook has more than twice as many figures as the earlier work, though about one-fifth are taken from another book by the reviser. While the number of pages is about the same as heretofore, reduction of certain appendixes of the earlier work made possible expansion of the text matter.

Typographical errors are few, as are errors of fact. Among the latter are the placement of the type locality of laccoliths in Nevada instead of Utah (p. 26) and a confusion of shape with roundness (p. 135). The reviewer would like a more complete treatment of tuffs, which are men-

tioned only in passing in the chapter on "General Remarks on the Igneous Rocks," whereas glasses warranted a three-page section of their own. In spite of these minor errors or shortcomings, the book is a worthy successor to its justly popular predecessor and will find a place in the library of many a student of rocks.

F. J. PETTIJOHN

Notice sur les travaux scientifiques de M. Emm. de Margerie. Macon: Protat Frères, 1938. Pp. 216.

Presented in this volume are the activities and accomplishments of an eminent geologist whose interests have been unusually international in scope. Anyone perusing the list of his publications will be struck by the very numerous expository and commentarial notes stimulated by new advances in knowledge in many portions of the globe. In a way they give the keynote of some of the chapters which follow. For these, while autobiographic in form and written from that point of view, at the same time bring out a wealth of general facts of significance in the growth of our science. One will naturally read with interest the parts played by Dr. de Margerie in these developments, but one will also be conscious that he is viewing a half-century of geologic progress delineated by a historian.

Separate chapters cover studies in the Pyrenees, the Jura, the north and east of France, geology of Europe, Asiatic studies, and the international geologic map of Africa, while "American Studies" makes the longest chapter of the volume. Nineteen chapters in all present the various lines of scientific work in which this versatile geologist has been engaged and of which he gives a valuable historical account.

R. T. C.

Géologie appliquée de l'Algérie. By M. DALLONI. Paris: Masson et Cie, 1939. Pp. 888; pls. 8.

As the title implies, this volume is devoted principally to the natural resources, especially the minerals, waters, and soils, and to the geology only as related to them.

Part I includes detailed discussions of the mineral resources, metals and nonmetals. Iron is the most important metal, followed by zinc and

lead. The iron occurs as magnetite and hematite in pre-Cambrian schists, as oölitic hematite in the Silurian and Devonian, as ferruginous conglomerate in the base of the Triassic, as pyrite in the top of the Triassic, and as deposits running as high as 54 per cent iron in Cenozoic oölites intercalated with argillaceous limonite. These deposits have achieved little commercial importance, owing to the lack of fuel in Algeria. There is some brown coal, mainly Cretaceous lignite. Some petroleum exploration work has been done, and the most favorable areas for prospecting are indicated. Part II deals with the subterranean waters and Part III with soils.

The Table of Contents is placed at the end of the book, and there is a general bibliography in the Introduction, with a bibliography on each subject following the treatment of that subject.

There is included a structural map of Algeria, a map showing the Triassic gypsum and other salt deposits, one showing the extent of the seas responsible for the phosphate deposits, and a map of the limestone areas.

LOUISE BARTON FREEMAN

Conservation in the United States. By A. F. GUSTAFSON, H. RIES, C. H. GUISE, and W. J. HAMILTON, JR. Ithaca, N.Y.: Comstock Publishing Co., Inc., 1939. Pp. 445; figs. 232. $3.00.

The purpose of this book is to present the basic facts essential to understanding the problems of conservation in a nontechnical manner for the use of students and the general reader interested in the question of depletion of the natural wealth of the United States. It is a symposium written by members of the faculty of Cornell University.

The Introduction and Part I, on the conservation of soil and water resources, were written by Gustafson, professor of soil technology. This part has four chapters which deal with the soil, its formation, the soil regions of the United States, and the productivity of the land; the use and conservation of water; soil depletion and erosion; and soil conservation, including flood control. Part II, written by Guise, former professor of forest management, deals with the conservation of forests, parks, and grazing lands. Part III, by Hamilton, assistant professor of zoölogy, treats of the conservation of wildlife. Part IV, by Ries, professor of geology, considers the conservation of mineral resources such as the metals, coal, petroleum and natural gas, and the nonmetallic minerals.

This excellent volume is fully indexed and includes an appendix of selections for supplementary reading. It is so arranged with questions at the end of each chapter that it can be used very well as a practical textbook for courses in conservation.

LOUISE BARTON FREEMAN

Soil Conservation. By HUGH HAMMOND BENNETT. New York: McGraw-Hill Book Co., Inc., 1939. Pp. 993; figs. 358. $6.00

Dr. Bennett is chief of the Soil Conservation Service of the United States Department of Agriculture and is one of the foremost authorities in his field. This book gives a detailed treatment covering every aspect of the subject of land loss due to soil erosion as well as its broad implications of social and economic decline.

The first part deals with soil erosion and includes a chapter on the beginning of erosion history with civilization in the countries of the Near East, from which it progressed westward into Europe and to the Americas. The results, processes, and rates of erosion are discussed and illustrated, and chapters on sedimentation and the relation of geology to soil erosion are included.

The second part is a consideration of soil conservation. Measures of soil defense by means of national-conservation action are discussed, and the need for a co-ordinated program for wise land use and treatment is emphasized. A chapter is given to each of the physiographic regions of the United States and their particular problems and to the relation of conservation to flood control. The volume is fully indexed and very well illustrated.

LOUISE BARTON FREEMAN